5. QUALITATIVE ENVIRONMENTAL DESCRIPTION

5.1 Atmospheric and Ocean Flows

Both the surface winds over the ocean and the upper ocean currents give rise to forces on the MOB as noted in Section 4. Because the surface winds are a major driving force for the ocean surface currents, an integrated discussion of atmospheric and ocean flows is presented. When transiting the oceans and when operating in different locations, the local winds and surface currents will, in the mean, differ as the site may lie in different climatological regions. For example, at a site off the west coast of Africa, between the Azores and Madeira, moderate trade winds out of the northeast at about 7 m/sec and about 50 cm/sec of surface flow in the ocean to the east associated with the Azores current would be anticipated.

To present an overview of the mean surface winds and currents to be expected around the world, the large-scale, climatological mean, general circulation of the atmosphere and ocean are presented first in Section 5.1.1. Then, Section 5.1.2 describes surface winds and related current that are regional, local, and transient; these winds and currents are often more energetic than the mean conditions and thus of more consequence to the design and operation of the MOB. Finally, Sections 5.1.3 and 5.1.4 describe additional regional and local surface currents to be found in the ocean that are not linked to local meteorology.

5.1.1 General Atmospheric Circulations and Associated Ocean Flows

The general circulation of the atmosphere at the ocean's surface is well known. The earth is heated more by the sun at the equator than at the poles. Surface heat and moisture fluxes drive convection, which together with the earth's rotation, gives rise to three meridional zones of surface winds. Equatorward of the polar highs are the bands of polar easterlies, which extend down to approximately 60° latitude. Equatorward of 60° are the westerlies, which extend down to roughly 30°. On either side of the Intertropical Convergence Zone (ITCZ) at the equator are the Trades. The relative thermal effects of the land distort this general pattern, leading in the annual mean to large scale gyre-like wind patterns over the ocean basis and to northern summer (July) and winter (January) mean wind fields that show clear evidence of monsoonal variability. The annual mean, the July monthly mean, and the January monthly mean surface winds from Tomczak and Godfrey's (1994) figure based on Trenberth et al's (1989) examination of winds from ECMWF (European Centre for Medium Range Weather Forecasts) are shown in Figures 5.1.1-1 through 5.1.1-3. There are a number of disturbances to the mean wind field which on regional (basin scale, such as the Arabian Sea), local (10s to 100s of km), and small scales (1 km) contribute variability which is often much more energetic than the mean. These disturbances will be discussed further below.

To a large extent, the surface circulation of the ocean is also well-known and shows a clear relation to the surface wind field. Figure 5.1.1-4 shows a map of the major surface currents from McLellan (1965). Note that large, basin-scale gyres in the surface currents roughly match the location of the gyres in the surface wind fields. A significant component of the surface current is driven by the local wind, accelerated by the surface stress the wind exerts on the sea surface. The Coriolis force associated with the earth's rotation causes the surface currents to be directed 45° to the right (left) of the surface wind in the northern (southern) hemisphere, and the subsurface currents turn further off from the wind with depth in the same direction (Figure 5.1.1-5).

The vertical integral of the Ekman spiral yields the wind-driven or Ekman transport, which is off to the right of the wind in the northern hemisphere and off to the left of the wind in the southern hemisphere. Thus, a combined plot of surface winds and wind-driven transport should show a clear relation. Figure 5.1.1-6 shows the surface Ekman transport at select locations on a map of the surface winds in the eastern Pacific.

The vertical integral of the wind-driven flow or Ekman transport is always expected to be 90° off the wind and has a magnitude equal to $\tau/(\rho_w f_c)$, where τ is the magnitude of the wind stress (roughly equal to $\rho_a C_d U^2$, where ρ_a is the density of the air, between 1.2 and 1.3 kg m^{-3} at the surface, C_d is the drag coefficient which rises from about 1.5 x 10^{-3} at 5 m/sec to about 4 x 10^{-3} at 50 m/sec, and W is the wind speed (in m/sec)), ρ_w is the density of seawater, between 1022 and 1028 kg m^{-3} at the surface, and f_c is the Coriolis parameter (=2ω sin l where ω is the angular velocity of the earth, 7.458 x 10^{-5} s^{-1}, and l is the latitude in degrees). However, the vertical structure of the wind-driven flow is not often found to be spiral-like. Locally wind-driven flow is restricted to be within the surface mixed layer of the ocean, where the mixed layer is the well-stirred region of nearly uniform temperature found in the upper few to upper hundred meters, as shown in Figure 5.1.1-7. Often, the wind-driven flow has a vertical structure with the vertical shear concentrated at the base of the mixed layer in the region of rapid change in temperature with depth (the thermocline). Thus, the wind-driven flow is sometimes referred to as "slab-like".

The locally wind-driven flow is present on top of flow that is supported by horizontal gradients in the density of the ocean. The mean density driven flow is known as the geostrophic flow and is characterized by a balance of Coriolis and pressure gradient forces ($-f_c\ u = -1/r_w$ ¶ $p/$¶x , $f_c\ v = -1/r_w$ ¶$p/$¶y). The mean density distribution in the ocean is indicated in Figure 5.1.1-8, which is a map of dynamic height at the surface relative to 2,000 m.

Dynamic height ($m^2\ s^{-2}$) provides an indication of the horizontal differences in vertically integrated density that support the pressure gradients, and geostrophic flow will be along its isolines, perpendicular to the pressure gradients.

The surface currents of marginal and enclosed seas are not as readily deduced from large scale forcing fields and have circulation features than reflect local dynamics. Figures 5.1.1-9 and 5.1.1-10 show the surface flow in the Mediterranean Sea and Sea of Japan as examples.

In the Mediterranean, topographic restrictions, including shoaling at sills, cause locally strong flows, up to 2 to 3 m/sec. Atlantic Ocean water flows into the Mediterranean Sea at the Straits of Gibraltar to replace the loss to evaporation within the Mediterranean;

speeds there reach 1 m/sec. The Sea of Japan also has limited access, but there is evidence that some of the water in the western boundary current, the Kuroshio, enters the strait between Japan and Korea and then exits in the straits in northern Japan.

5.1.2 Secondary Atmospheric Circulations and Associated Ocean Flows

Embedded within the mean surface wind field are a number of secondary circulations that locally may be very strong and dominate the surface wind variability for periods of months to minutes. The time and space scales of these secondary atmospheric flows are, in general, linked as indicated in Figure 5.1.2-1. At many ocean locations, there is more variability and energy associated with synoptic weather events (5- to 20-day periods) than with either lower frequency (i.e., mean) or higher frequency winds. The winds associated with these various phenomena give rise to surface currents. Both are discussed here.

To a large extent, in the distribution of energy in both the surface winds and the surface currents, the lower frequencies have more energy. Figure 5.1.2-2 (from Brink, Moyer, Trask and Weller (1995)) shows the autospectra based on two years of surface meteorological observations from a buoy at 33° N, 22° W in the eastern North Atlantic. Note that the spectrum of the wind stress slopes upward, going from high frequencies to lower, until about 0.01 cph (cycles per hour). At 0.01 cph and lower, the spectral levels are flat. However, this is a trade wind region, with relatively low levels of cyclogenesis and synoptic weather variability associated with atmospheric fronts and lows. In many other locations, the wind stress spectra show a broad peak at periods of 0.01 to 0.002 cph (roughly 5- to 29-day periods) associated with synoptic weather variability. Figure 5.1.2-3 (from Brink, Moyer, Trask and Weller (1995)) shows the autospectra of ocean currents at 10-, 90-, and 200-m depths at the same site. In these plots, the 10-m spectra are in correct position, the 90-m spectra are shifted down 2 decades from the 10-m spectra and the 200-m spectra are shifted down 4 decades from 10-m spectra. Note spectral peaks at the local inertial (f) and tidal (M_2) frequencies and the continuous increase in energy with lower frequencies. The ocean spectra do not show the loss of energy at low frequencies, as low period ocean eddies are energetic at this site. In both cases, spectra are a representation of the average energy level at each frequency and do not indicate whether there were a few very strong events or many moderate events contributing to the variability at a given frequency. However, such information can often be found by researching the variability specific to a given site or path for transit.

5.1.2.1 Monsoons

Relatively large-scale circulations are found that have strong seasonal cycles. Seasonal differences in heating the land masses and resulting gradients in barometric pressure give rise to monsoonal wind patterns in the Indian Ocean, over Southeast Asia, and to a lesser extent over the Americas. The monsoon surface winds are striking in the steadiness of direction and persistence, lasting months; they thus can result in significant wind-driven currents and wind waves. Figure 5.1.2.1-1 illustrates the winds and currents associated with the Indian monsoon. The summer, Southwest Monsoon, is notable because the highlands in Kenya block the flow of the monsoon winds over Africa, turn and concentrate the winds into strong (monthly mean of 14 to 15 m/sec) winds flowing parallel to the Yemeni-Omani coast. These winds are referred to as the Somali Jet or the Findlater Jet. The northern Indian Ocean monsoon is also notable because, in this region, the western boundary currents reverse direction with the winds. Monsoonal variability in winds and wind-driven currents is found in other Asian/Indonesian Seas. Figure 5.1.2.1-2 illustrates the monsoonal variability of surface currents in the South China Sea.

Monsoons are of note because of their duration, strength, and steadiness. In the Arabian Sea, the Southwest Monsoon, once blowing, exhibits a wind direction that does not deviate 5°, and though the strength varies, this unidirectional wind may persist from June through December. The speed may be 15 m/sec, and with a long fetch, significant waves may develop. Prediction of the exact time of the onset is not yet possible. From year to year, the time of the onset varies and may be abrupt or turn on briefly, then off, and back on (multiple onset).

Monsoonal winds drive surface currents, as do other winds. Thus, reversal of the surface currents may also be expected. Monsoon winds blowing parallel to a coast (with the coast to the left in the northern hemisphere) can develop strong upwelling regimes, bringing cool water up from below, with embedded jet-like flows that carry the cool water far offshore. In addition, large eddies may be formed, as in the lee of Socotra Island in the Arabian Sea.

The steadiness, strength and repeatability of monsoonal winds lead to the simplest conditions for wave generation and growth. Under these conditions, simple steady-state models, such as those described in 5.3.4.1, are adequate to the task of hindcasting wind-sea in areas of weak currents. However, the very persistence of monsoonal wind systems leads to the development of significant oceanic momentum, which, through Coriolis effects, are often concentrated in strong western boundary currents such as the Somalia current and the Agulhas current. Coastal jets and other strong current systems may have pronounced effects on the local wave climate through wave-current interactions. These interactions may be both kinematic, leading to refractive effects and focusing and defocusing of wave energy, and dynamic, producing locally large variability in wave heights. The focusing and modulation of waves in the Agulhas current have been blamed for the appearance of "freak" (or giant) waves and the consequent extensive damage to ships.

The horizontal scale of these wave-current interaction processes and their effects cover a wide range. Strong local modulation of wave heights due to horizontal current shears - waves steepen as they approach an increasing adverse current - may be on a scale commensurate with that of the strong gradients on the edges of current systems. These may be of the order of 1 km, i.e., much smaller than the grid scale of typical basin-wide spectral wave prediction models. On the other hand, refractive effects of currents, which can be unsteady as well as inhomogeneous, may produce focusing of wave energy at substantial distances from the locale of the wave-current interaction. Generally, finite difference models are not well suited to handle these effects on a basin-wide scale. It may be necessary to have a hierarchy of models nested in such a way as to improve the resolution in areas of strong current gradients. In addition, ray tracing of long swells over bottom bathymetric variations and through current shears may be necessary to localize areas of focused wave energy.

5.1.2.2 Tropical Cyclones/Hurricanes

Cyclones that originate over tropical waters are called tropical cyclones that, when strong enough, are also called hurricanes or typhoons. They can be very intense, with winds of up to 90 m/sec. They have a well-defined radial structure with winds revolving around a central low as shown in Figures 5.1.2.2-1 and 5.1.2.2-2.

The large size, intensity, and potential damage upon landfall of tropical storms have led to considerable attention in tracking, observing, and predicting hurricanes. They are, in general, well-represented by the analyses produced by the weather forecast centers and skill at predicting their tracks is growing. In addition, there are good records about when and where tropical storms and hurricanes are likely, including compilations of storm tracks that could be examined as a basis for estimating the likelihood of a hurricane occurring in a given location. Tropical cyclones should be anticipated off the U. S. East Coast, in the Gulf of Mexico, off Baja California, in the Arabian Sea and Bay of Bengal, in the South Indian Ocean, in the Yellow Sea, from Indonesia up through the Philippines, the South China Sea, the Sea of Japan, and east of Japan, and in the South Pacific east of Australia.

If hurricanes move quickly (i.e., do not spend a significant fraction of an inertial period, $12/\sin\theta$ hours, where θ is the latitude) over the ocean, they have little effect on currents, and thus give rise to little signature in the surface currents. Typically, however, there are two principal effects that each give rise to surface currents. First, the winds rotating around the low pressure center, transport surface water radially outward (Ekman transport). This results in intense upwelling along the track of the hurricane that results in a cold ridge (Figure 5.1.2.2-3) in the upper ocean thermal structure. The resulting density gradients support geostrophic flows as strong as 50 cm/sec. In addition, strong oscillatory currents that change direction 360° over the inertial period are also generated; these are referred to as inertial oscillations and may have speeds of up to 100 cm/sec. An example of the generation of inertial oscillations is shown in the next section.

These are the strongest wind systems of a scale large enough in space and time to cause the growth of waves sufficient to affect a MOB. In rough terms, the requisite space/time scales to provide an approximate doubling of wave energy at a given period, T, is about 1000 times T; the space scale is set by the distance of energy propagation (group speed) over that time. For a wave of 10-second period, which can be 20 m high, these scales are about 3 hours and 80 km. Hurricanes have typical time scales of several days and space scales of a few 100 km. In terms of efficiency of wave generation, a stationary hurricane is over-endowed with time relative to fetch and so the waves emanating radially from the area of strong winds are essentially fetch limited. When hurricanes travel, the waves to the right (left) of the storm in the northern (southern) hemisphere are moving in the same direction as the winds that are forcing their growth. In this way, the hurricane increases the wave generating fetch for these waves and, in the fortuitous circumstance of rate of travel in a straight line at the group velocity of the largest waves in the system, may produce exceedingly large waves in the area of the storm where tracking direction and wind velocity are parallel. This additional degree of freedom of hurricanes makes the estimation of extremes from such storms very difficult to deduce. In principle, if a hurricane maintains course (more or less, without rapid changes) at the optimal group velocity over 5000 km and the tangential wind speed averaged radially over 100 km is 50 m/s, it could produce waves with peak periods of 27 seconds and significant heights of 40 m. To do this it would have to be traveling at 21 m/s at the end, and such tracking speeds are unusually high. Nonetheless, it is clear that hindcasting waves in hurricane prone areas can only achieve some accuracy if the tracking characteristics can be specified.

As explained above, wave current interaction is not as important here as it is for monsoonal areas because fast moving hurricanes do not spin up substantial currents and their passage over strong boundary currents is too rapid to produce a sustained effect. However, some caution should be exercised here, because it is possible that the steepening or focusing of a raging sea beneath a tracking hurricane, albeit transient, may produce a combination of extremely high waves and strong winds that will test the mettle of any craft.

5.1.2.3 Extra-tropical Cyclones and Other Large-Scale Storms

Extra-tropical cyclones develop along atmospheric fronts found in mid and high latitudes and are thus distinct from tropical cyclones by their formation region. Figures 5.1.2.3-1 and 5.1.2.3-2 show the evolution of a cyclone spawned along a front in the Yellow Sea in November 1972, and the mean storm tracks for that region. The observed wind speeds associated with this cyclone reached 23 m/sec.

Some extra-tropical storms can be small in scale (less than 100 km wide) and intensify very rapidly. Because of this, they are not resolved well by numerical weather prediction models, which have grid scale of 100 km and larger and time steps of 6 hours. In late 1978, for example, a small, strong cyclone spawned in the North Atlantic caught the Queen Elizabeth II unaware and damaged the ship. Figure 5.1.2.3-3 shows a short, intense wind event observed in the Arabian Sea off Oman that was as energetic as the monsoon but was not predicted in forecasts received by the ship at the site or seen in later analyses.

Storms passing over the ocean have two primary effects on the upper ocean. They mix the water and cause the upper, mixed layer of the ocean to become deeper. They also impulsively add momentum. The ocean responds to this impulsive forcing at its resonant frequency, which is the inertial frequency $f_I = (\sin l) / 12$ in cycles per hour. The mixed layer moves off like a slab and executes a circular trajectory. As a result, the velocities at a fixed point turn through 360°. Figure 5.1.2.3-4 shows current meter records from the upper ocean as several storm events pass the location. On approximately day 275, the wind forcing generated strong inertial oscillations with amplitudes close to 50 cm/sec. Note, though the details of the wind forcing are important to exciting these oscillations and other wind events, such as on day 259, they were not as effective at exciting inertial oscillations. In addition to inertial oscillations, local storms would force surface currents of lower magnitudes at lower frequencies. The surface mixed layer would absorb momentum from the wind field and move off relative to the ocean below. Note also that inertial oscillations propagate down from the mixed layer and can be found in the thermocline and that there is a phase shift across the thermocline. Thus, a MOB hull that is deeper than the mixed layer may see some vertical shear at the inertial frequency.

The rapid intensification ("explosive cyclogenesis") of extra tropical storms is the characteristic of most interest to the estimation of wave properties in these areas. Here, as for hurricanes, their tracking characteristics are important. In addition, their sudden growth and the frequent occurrence of more than one system in an ocean basin means that confused seas (having multi-modal directionality) and the interaction of crossing wave trains may have to be considered.

5.1.2.4 Local Winds

Locally, wind variability will be experienced from additional sources. In some cases, that wind will also lead to surface currents.

Atmospheric fronts and squall lines can pass over a location, bringing abrupt changes in wind speed and direction. The passage of convective systems overhead will be experienced at the surface as rain, with winds reversing direction and strengthening as the radially outward flowing surface winds below the downdraft region are felt.

Along and near coasts, mountain forms can channel winds and intensify them. West of Central America, for example, the prevailing trade winds are funneled through mountain gaps and produce strong local winds that blow over the Gulf of Tehuantepec. These winds are known as Tehuantepecers. Similarly, strong Santa Anna winds can be channeled by canyons and blow in intense, local jets off the coast of California. Local mountains can block, channel, and thus intensify winds parallel to the coast as well. The Southwest Monsoon in the Arabian Sea was discussed earlier. On a smaller scale, similar steering and intensification has been seen on the southwest coast of Greenland where northeasterly winds can reach hurricane force and in western Washington state where local winds can exceed 51 m/sec in what has been called a mesoscale cyclone induced by the shape of the mountains.

Winds flowing parallel to the coast can give rise to offshore currents. The wind accelerates the surface currents but they are deflected by the Coriolis force to the right (left) in the northern (southern) hemisphere. For example, southward winds along the U. S. west coast give rise to wind-driven flow directed offshore. This water is replaced by water that upwells from below. Thus, the surface currents off California are a combination of eastern boundary currents (see below, Section 5.1.3) flowing parallel to the coast and wind-driven flow carrying water offshore.

Low-level wind jets exist in some coastal locations. One is seen in the Gulf of Mexico. The maximum wind speeds (up to about 15 m/sec) are typically at 200- to 300-m altitude, with reduced winds at the surface. Near shore, a daily reversal in winds can be experienced in association with sea breezes. The thermal effect of western boundary currents such as the Gulf Stream may also give rise to local sea-breeze-like wind patterns. There may be daily variability in the wind away from land as well if the change in the air-sea heat flux alters the vertical structure and/or stability of the atmospheric boundary layer. In stable conditions, the shear on the surface wind is concentrated near the surface, while in unstable conditions, the wind in the deeper mixed layer is less sheared.

These wind features will have associated wind-driven surface currents whenever the wind blows for longer than the inertial period. Reversing and transient winds, though strong, will excite little wind-driven surface current if they last less than an inertial period in the open ocean. Thus, though a squall line passing over the MOB in 20-minutes time may have strong, gusty winds, it will not lead to surface currents of consequence. A persistent wind, like a Tehuantepecer, will, on the other hand, drive local wind-driven flow and create surface waves.

Under the label of local winds we include squall lines, land-sea breezes and orographically forced flows. Squall lines can be very sharp with strong spatial wind gradients. They have been known to propagate at propitious speeds for optimal wave generation, but they are too short lived to generate waves large enough to affect the MOB. Land-sea breezes are generally too weak and confined (typical extents are no more than 50 km from the coast) to account for waves of any significance in the MOB world. Orographically intensified flows, such as the Tehuanapecer and the Santa Anna winds, can be very strong and persistent and can therefore generate rapid wave growth. However their intensity drops off away from the coastline giving relatively short fetches for wave growth. These wave growth conditions are easily modeled with fine resolution spectral models and, inasmuch as the strongest flows are offshore, an adequate job can be done with steady-state significant height, period and direction models such as those described in Section 5.3.4.1.

5.1.3 Boundary Currents, Eddies, Jets, and Fronts in the Ocean

In some locations, the geostrophic (driven by the density gradients rather than by the wind) flows in the ocean are a significant and at times major component of the surface flow. Along the western boundary of the ocean basins, as a consequence of general westward drift of energy in the oceans, persistent strong western boundary currents are found. These may have flows in excess of 250 cm/sec. Examples are the Gulf Stream of the eastern United States (Figure 5.1.3-1), and the Kuroshio and Oyashio off the eastern coast of Japan (Figure 5.1.3-2). For part of their flow along the coast, these currents may be stable and essentially jet-like features. As the boundary currents separate from the coast and turn back into the ocean, they are unstable. They meander, and their location varies with time. They also fold, pinch off, and thus generate eddies.

The strong boundary currents of the Indian Ocean are seen in Figure 5.1.3-3. In this figure (and Figure 5.1.3-4) wavy lines show ocean fronts (boundaries between regions with different water properties). Of particular interest in the Indian Ocean are the boundary currents of Somalia that reverse with the monsoon and the extension of the Agulhas

Current south of South Africa where it bends back and returns eastward. This bending back has been called the Agulhas retroflection and leads to a locally complex system of surface currents.

The Pacific Ocean (Figure 5.1.3-4) also has western boundary currents, off Australia and Asia. Note also the strong zonal currents near the equator and around the Antarctic.

The strong western boundary currents give rise to eddies. These form as convolutions of the western boundary current, but can be pinched off and drift away (both toward shore and toward the open ocean) as isolated features as shown in Figure 5.1.3-5. Such eddies have a typical scale of 100 km and velocities of up to 1 m/sec. Eddies shed by the Gulf Stream have been called Cold Core Rings (cold water in the center) and Warm Core Rings (warm water in the middle).

The Caribbean Current that flows northwest between Cuba and Central America enters the Gulf of Mexico where it forms the Loop Current (Figure 5.1.3-6). There has been considerable effort devoted to observing the Loop Current because it moves and also sheds energetic eddies that affect oil drilling operations in the Gulf of Mexico.

A feature of boundary regions is that the flow may not be found to be stationary and uniform. It may be transient and variable in space. Figure 5.1.3-7 illustrates the flow along the coast of Somalia showing two gyres and a strong, offshore jet. Such jet-like features are sometimes locked to topography and extend offshore from capes. They have been called coastal jets and sometimes coastal squirts, may be narrow (10 km across) and may extend 100 km offshore.

Away from ocean boundaries, rather strong local currents can be encountered in association with rapid transitions in water properties. In analogy to similar atmospheric features, these are called ocean fronts. The advection of an open ocean front past an array of five surface moorings is shown in Figure 5.1.3-8. The along-front velocities of the frontal jet fit within the array, which was approximately 50 km across the longest side. In a vessel, steaming across a front, water temperature may change several °C and an along-front current of up to 1 m/sec may be encountered that has a sharp boundary and is about 30 km in width. The front may resemble a rather stable, jet-like feature. It may also be unstable, giving rise to convolutions and eddies. Small, intense eddies have been reported in association with fronts by Weller and Samelson (1991) and are shown in Figure 5.1.3-9. These eddies with widths estimated to be 12 to 20 km had speeds as high as 60 cm/sec.

5.1.4 Equatorial Currents

Typically, equatorial currents away from the boundaries are zonal. Figure 5.1.4-1 shows the surface currents in the equatorial Atlantic. Seasonal variability is evident, as are the strong boundary currents along the east coast of South America. Temporal variability of the equatorial currents is even more pronounced in the Indian Ocean (Figure 5.1.4-2). The Equatorial Jet that appears in April through June may be as strong as 70 cm/sec. The North Equatorial Current (NEC) runs at about 30 cm/sec from the Malacca Strait but strengthens to 50 to 80 cm/sec between 60° E and 75° E. The South Equatorial Current

(SEC) has speeds of 30 cm/sec and less. The Equatorial Countercurrent (ECC) flows east in the northern winter with speeds of between 30 and 80 cm/sec. This figure also illustrates the variability of the monsoon-related surface currents. The East Indian Current (EIC) present in northern spring east of Indian is replaced in the northern fall by the East India Winter Jet, which flows in the opposite direction.

5.1.5 Other Currents

A final type of ocean phenomenon that will be described here is an internal wave, which results in significant surface currents.

The density stratification of the ocean interior enables the existence of a class of motions (termed "baroclinic") which are manifest on the sea surface primarily in terms of their horizontal currents. Conventional ships notice few aspects of these sub-surface flows, aside from the low frequency ($f < 0.1$ cph, typically) lateral movement of the surface. The MOB in contrast, draws sufficient water that aspects of the structure might well extend into the stratified region of the upper ocean. This has implications for both the structural design and operation of the various MOB subsystems.

5.2 Phenomenological Description of Environmental Components

5.2.1 Wind

Descriptions of the wind flow typically required for structural engineering purposes are provided by:

(1) Micrometeorological models - spatial and temporal models of the wind flow. These models are normalized with respect to the mean wind speed at a reference elevation (e.g., 10 m above the Earth's surface).

(2) Wind climatological models - models of the probabilistic behavior of extreme wind speeds at the reference elevation, for any given direction or regardless of direction. Such models may, in principle, be used to estimate extreme wind speeds for average return periods varying from periods of the order of 10 years up to as many as 10,000 years, though sampling errors for long return periods are typically very large for individual records, which usually contain no more than a few tens of data points. In some instances such sampling errors can be reduced by techniques that use spatial information on extreme speeds.

5.2.1.1 Wind Measurements

Wind speed measurements are needed for:

1. Estimating extreme wind speeds near the ocean surface. Measurements of extreme speeds are rarely obtained for destructive hurricanes (we use this generic term also for typhoons and Australia/Indian Ocean region cyclones), since wind measurement

systems invariably fail in extreme events. In the absence of measurements of extreme hurricane winds, measurements of quantities such as the largest atmospheric pressure deficit, the radius of maximum wind speeds, hurricane translation velocity, and other relevant quantities can be used to make indirect inferences on maximum hurricane wind speeds and directions for design and hazard assessment purposes.

2. Modeling as correctly as possible the structure of the wind flow near the Earth's surface, that is, the mean wind speed profile and the temporal and spatial variation of the wind speed and direction. Records of measurements that may be used to construct or verify models of the wind flow structure are available for large-scale extra-tropical storms, but only in very few instances for strong hurricane winds (i.e., winds near the eye) and for smaller-scale storms (e.g., local thunderstorms). Measurement systems for tornado winds are only now beginning to be developed.

Note, in particular, that the absence of sufficient data on hurricane wind profiles makes it virtually impossible to convert reliably information on hurricane winds near the eye measured at high altitudes to information on wind speeds near the surface.

Time series of winds are available over the ocean at select locations that could be used to develop statistical descriptors and also as a source of time series for simulations. For some years, Ocean Weather Stations (OWS) were maintained in the North Pacific and the North Atlantic. Only one, OWS Mike, is still in operation by the Norwegians. These records are valuable due to the strong winds (observations every 3 hours) and because they come from locations with little other data. OWS Papa (50° N, 145° W) and OWS November (30° N, 140° W) data have been used for the development of numerical models of upper ocean variability (see, for example, Martin (1985)).

From the mid-1980s research groups have mounted campaigns with surface moorings at select locations. The Upper Ocean Processes Group at the Woods Hole Oceanographic Institution has wind time series sampled typically every 15 minutes from the locations over the times indicated in Table 5.2.1.1-1.

Experiment Name	Location	Dates of Buoy Deployment
Joint Air-Sea Interaction Experiment (JASIN)	59° N, 12.5° W	07/1978 - 09/1978
Long-Term Upper Ocean Study (LOTUS)	34° N, 70° W	05/1982-05/1984
Frontal Air-Sea Interaction Study (FASINEX)	27° N, 70° W 5 mooring array	01/1986 - 06/1986
BIOWATT	34° N, 70° W	05/1987-11/1987
Severe Environment Surface Mooring (SESMOOR)	42.5° N, 61.2° W	10/1988 - 03/1989
Mixed Layer-Marine Light (MLML-89)	59.5° N, 20.8° W	04/1989 - 09/1989
Surface Wave Process Program (SWAPP)	34° N, 127° W	02/1990 - 03/1990
Surface Wave Dynamics Experiment (SWADE)	34° N, 68° W	10/1990 - 04/1991
Acoustic Surface Reverberation Experiment ASREX 91)	49° N, 131° W	10/1991 - 01/1992
MLML 91	59.5° N, 20.8° W	04/1991 - 08/1991
Subduction	18°-33° N, 22°-33° W 5 mooring array	06/1991 - 06/1993
TOGA Coupled Ocean-Atmosphere Response Experiment (COARE)	1.75° S, 156° E	10/1992 - 03/1993
ASREX III	33.9° N, 69.7° W	12/1993 - 03/1994
Arabian Sea	15.5° N, 61.5° E	10/1994 - 10/1995
Pan American Climate Study (PACS)	3° S, 125° W 10° N, 125° W	04/1997 - 09/1998

Table 5.2.1.1-1. Locations and Durations of WHOI Surface Mooring Deployments. These provide wind and upper ocean velocity records.

Beginning in 1975, NOAA established an array of moorings (called the TAO array) in the tropical Pacific, between 8° N and 8° S that provide daily winds from this region; this data is available through the TAO homepage on the web. The National Data Buoy Office also can provide the records from the coastal weather buoys maintained around the United States.

5.2.1.2 Horizontal Variations and Coherence

Moored array experiments have examined spatial variability in the ocean wind field at scales down to approximately 15 km. Spatial variability at these scales comes from small-scale atmospheric systems (lows, fronts). Comparisons of buoy and ship observations from the tropics show somewhat smaller scales associated with convective systems (downdrafts and squall lines). Mesoscale atmospheric modeling and research aircraft flights in the atmospheric boundary layer at a height of 30 m above the sea surface provide a source of data, but resolve only down to about a 1-km horizontal scale.

5.2.1.3 Models of Flow Structure near the Surface

For structural engineering purposes the flow structure near the surface is defined by information on mean wind speed profiles, turbulence intensity, integral scales of turbulence, turbulence spectra, and turbulence spatial cross-spectra. Such information is available mostly for large-scale extra-tropical storms and, as indicated subsequently, only in very few instances for other types of storms. In particular, virtually nothing is known about the actual flow structure near the surface in tornadoes or water spouts.

Considerable differences exist in the character of winds from different sources. Figure 5.2.1.3-1 shows a wind record from a large-scale extra-tropical storm over a little less than 2 hours. The mean value varies little during this period, but considerable turbulence is seen. Figures 5.2.1.3-3 cover about 2½ hours of a thunderstorm. The very rapid rise in wind speed (accompanied by a similarly rapid change in direction) as the front passes is seen (the plot runs from right to left). Hurricanes have change of speed and direction intermediate between the above two storm types. Figure 5.2.1.3-2 shows wind speed during a hurricane, over a 4-hour period, with the eye passing over the site at about 11:30. All the above figures are from Simiu and Scanlan (1998).

Despite the differences in the character of the wind, the same models of turbulent velocity fluctuations are currently used for all types of storms, with the exception indicated in the section describing the relation between mean wind speeds and speeds corresponding to various averaging periods. In the absence of better information, all models described subsequently are commonly assumed to be valid for ocean storms as well as for storms over open land.

<u>Mean Wind Profiles</u>

Mean wind profiles in large-scale extra-tropical storms are described by the logarithmic law

$$U(z)=(1/k)u^*\ln(z/z_o) \qquad (5.2\text{-}1)$$

where U denotes the mean wind speed averaged over about 10 minutes to one hour, $k \approx 0.4$ is the von Kàrmàn constant, z is the elevation above the zero plane displacement z_d, u^* is referred to as the friction velocity, and z_o is the roughness length. For flows over open terrain and the ocean surface, it is customary to assume a zero plane displacement $z_d = 0$ and roughness lengths $z_o = 0.05$ m and $z_o = 0.005$ m, respectively. However, for strong flow over water, z_d and z_o are in fact not clearly defined. Wind speed dependent models of the roughness length for flow over water have been proposed - see Simiu and Scanlan (1996), p. 45 and references therein. According to these models, for wind speeds 4 m/sec $< U(10) \leq 20$ m/sec,

$$z_o = 10 \exp\{-17.7\,[U(10)]^{-0.23}\} \qquad (5.2\text{-}2)$$

or

$$z_o = 10 \exp\{-40[7.5+0.67U(10)]^{-1/2}\,]\} \qquad (5.2\text{-}3)$$

(in meters), where U(10) is the mean wind speed in m/sec at 10 m above the mean water level. For example, for U(10) = 20 m/sec, z_o = 1.38 x 10^{-3} m according to Eq.5.2-2, and z_o = 1.59 x 10^{-3} m according to Eq. 5.2-3. For U(10) > 20 m/sec or so it has been suggested that z_o is constant.

The logarithmic law also appears to be valid for thunderstorms, even though the latter are not boundary-layer winds properly so called. It is also valid in hurricanes sufficiently far from the eyewall (i.e., from the region of maximum wind speeds). Its validity - or otherwise - for hurricane winds near the eye remains to be established. An alternative model of the mean speed profile is the power law, which is still used in some structural engineering standards - see Simiu and Scanlan (1996), p. 46.

For relatively low wind speeds, the flow stratification is, in general, not neutral. (Neutral stratification is the condition prevailing in the atmosphere wherein the temperature difference between two levels is the same as the temperature difference corresponding to an adiabatic lapse rate, i.e., as the temperature difference associated with the respective atmospheric pressures.) In such flow, temperature effects can be significant, and the mean speed profile can deviate significantly from the logarithmic law -- see Simiu and Scanlan (1996), pp. 49-51. According to estimates in Simiu and Scanlan (1996), the deviations from the logarithmic law tend to be significant (i.e., about -15% at 15 m elevation) for U(10) ≈ 5 m/sec, but decrease fairly rapidly for larger speeds.

Of special interest to the MOB, the spatial variation of mean wind speeds at any given time can be significant over a 1 km scale for local storms of the downdraft type, and, to a lesser extent, for hurricane wind speeds. In the latter case, models of that variation are inherent in modeling the translating hurricane vortex, which is discussed, e.g., in Simiu and Scanlan (1996), pp. 112-114. The smaller the diameter of the hurricane eye, the larger is the variation in mean wind direction on the scale of one kilometer.

Inferring wind speed profiles over the ocean near the coastline from wind speed profiles near the coastline over the ground requires modeling of the relationship between surface winds and gradient winds (i.e., winds at the top of the atmospheric boundary layer, where surface friction effects are negligible). For extra-tropical storms, gradient winds are usually modeled as geostrophic winds (i.e., the centripetal acceleration that affect the gradient winds is neglected). The modeling of the relationship between surface winds and geostrophic winds has a reasonably well established basis in theory and measurement. According to that relationship, in fully developed flow over water and over open terrain, the ratio of the respective mean wind speeds at 10-m elevation is about 1.2. If the flow is not fully developed, that is, if the distance between the shoreline and the anemometer at elevation h over open terrain is less than, roughly, 12.5 h, for winds blowing from the ocean that ratio is less than 1.2 (see Simiu and Scanlan (1996), pp. 71-73). This is shown in Figure 5.2.1.3-4. In region 1 (above line AB, approximately defined by a slope of 1:12.5), the wind speed is essentially equal to the wind speed upwind of the discontinuity. In region III (below the line AC, defined by a slope of about 1:100) it may be assumed, at least approximately, that the flow is determined by the same parameters u* and z_o that control the flow far downwind. Region II is a transition region.

For hurricanes, only tentative models exist of the ratio between surface winds and winds unaffected by surface friction. Hurricanes are vortex-like storms, with rotational wind speeds that depend upon the radius from the center of the hurricane eye. Those speeds increase up to a maximum radius R_m, referred to as the radius of maximum wind speeds, then decrease as the radius increases beyond R_m. In the absence of more appropriate models, it has been common practice in structural engineering applications to use empirical models based on judgment, simplified models, and some measurements. One such models is

$$U(10,R_m)=0.865\ V_{gr}\ (R_m)\ +0.5\ V_f \qquad (5.2\text{-}4)$$

where $U(10,R_m)$ is the wind speed at 10 m above the ocean surface, averaged over 10 min, R_m is the radius from the hurricane center to the circle with maximum wind speeds, V_{gr} is the calculated gradient wind, and V_f is the hurricane translation speed (see Simiu and Scanlan (1996), pp. 113 and 116 for details).

Turbulence Intensity and Integral Turbulence Scales

The turbulence intensity at a point is defined as the ratio of the r.m.s. of the instantaneous wind speed to the mean wind speed at that point. For wind flows over the ocean the longitudinal turbulence intensity (i.e., the intensity of the longitudinal component of the turbulent fluctuations) is approximately $1/\ln(z/z_o)$.

Integral turbulence scales are measures of the spatial dimensions of the turbulent eddies. Similarity with respect to turbulence intensity and turbulence scales is required, at least to some extent, for wind tunnel simulations of aerodynamic forces. For a summary of information on turbulence intensities and integral scales of turbulence, see Simiu and Scanlan (1996), pp. 52-55. The longitudinal turbulence scale can be very roughly described by the empirical formula

$$L_u^x = Cz^m \qquad (5.2\text{-}5)$$

which appears to yield scales that are larger (by factors of as much as 2) than measured scales. The constants C and m are shown in Figure 5.2.1.3-5 (from Counihan (1975)). The integral scales L_u^y and L_u^z (i.e., the measures of the lateral and vertical sizes of the turbulent eddies associated with the longitudinal component of the turbulent fluctuations) are, roughly, about one third and one half of the scale given by Eq. 5.2-5.

For a high mean wind speed (using $z_0 = 1.4 \times 10^{-3}$), we get $L_u^z =$ 0.2 miles at an elevation of 30 m (corresponding to the deck). Thus the scale of turbulence in winds during storms is about 1/5 of the length of a MOB, or about the length of one module.

For structures with nonlinear behavior, such as a MOB, the estimation of the turbulence effects requires in general the use of time-domain solutions wherein the atmospheric turbulence and its three-dimensional spatio-temporal properties are simulated numerically. Numerical simulations are also in order in situations when the phase relation between fluctuating responses of various parts of the structure is required. Computational Fluid Dynamics solutions of the Large Eddy Simulation type or other types may be used, especially if interest in higher-frequency, smaller scale turbulence components is secondary. Alternative types of modeling have been developed by Veers (1988) and Mann (1998). According to Mann, his procedure, used for load calculations on wind turbines and bridges, is more efficient and simpler to implement than Veers'. Mann's procedure is capable of taking into account the models chosen for the spectra and automatically yields the corresponding coherences. For a wind speed of 40 m/s, at an elevation z = 40 m, for horizontal separations of 40 m to 60 m, and for very low frequencies (e.g., n = 0.025 Hz), the use of simplified models of the coherence (Eq. 5.2-7) yields results comparable to those obtained by Mann (1998), that is, the Mann (1998) procedure does not necessarily result in reductions in coherence levels corresponding to large separations and small frequencies, whereas such reductions appear to be inherent in the models of Mann (1994) and Krenk (1996). The procedure of Mann (1998) is based on (1) the hypotheses of the rapid distortion theory of turbulence, entailing a linearization of the Navier-Stokes equations of motion of the flow, and (2) Taylor's hypothesis, according to which the turbulence is "frozen," i.e., transported with the mean velocity of the flow. The extent to which these hypotheses are applicable to eddies associated with very low frequencies – which are of primary concern in MOB design - needs to be examined, but as a crude approximation the procedure of Mann (1998) may be useful, provided that spectral models used in conjunction with it are appropriately chosen.

Relation Between Mean Wind Speeds and Speeds Corresponding to Various Averaging Periods

This relation is needed for structural engineering and wind climatological purposes. In particular, the relation between mean speeds (averaged, e.g., over one hour, or over ten minutes) and peak speeds (maximum speeds averaged over, say, two or three seconds) is of interest. Fairly well established models of such relationships are available for extra-tropical storms. More recently, based on a relatively small number of hurricane records, a tentative model has also been developed for hurricanes. Both models are described in the Commentary to the ASCE 7-95 Standard (1995) (Figure 5.2.1.3-6). See also Simiu and Scanlan (1996), pp. 69-70, 78-79.

Spectra and Spatial Cross-Spectra of Turbulent Wind Tunnel Fluctuations

Spectra of wind speed fluctuations are needed primarily for applications in which the dynamic amplification of effects associated with the oncoming longitudinal wind fluctuations is significant.

Reasonably reliable models exist for the turbulent components at relatively small periods, say, 5 seconds or less. However, for large periods, existing models, though reasonable, are to some extent tentative. This is true particularly of fluctuations with periods of the order of tens of seconds or longer. For a survey and details on spectra of the longitudinal, lateral and vertical wind speed fluctuations see Simiu and Scanlan (1996), pp. 59-64. For the purpose of designing structures with very low natural frequencies of vibration (say, less than 0.1 Hz), the spectral density of the longitudinal velocity fluctuations yielded by Eqs. 2.3.25 of Simiu and Scanlan (1996), pp. 62-63 may be considered. An example of spectral shape yielded by those equations for the particular case of flow over the ocean at 35-m elevation (U(35)=45 m/sec) is shown in Figure 5.2.1.3-7 (see Simiu and Scanlan (1996), p. 63).

Fluctuations at different points in space are not perfectly coherent, i.e., a peak speed at one point may not occur at the same time at another. The degree to which this is the case is described, for each individual fluctuation frequency, by the cross-spectral density function, for which a commonly used empirical expression is

$$S^{cr}(r,f) = S^{1/2}(z_1,f)S^{1/2}(z_2,f)\ \exp\{2f[a^2\,(y_1-y_2)^2 + b^2\,(z_1-z_2)^2\,]/[U(z_1)+U(z_2)]\} \qquad (5.2\text{-}6)$$

where r is the distance between two points in a vertical plane normal to the mean wind velocity whose elevations and horizontal coordinates are denoted by z_1, z_2, and y_1, y_2, respectively, f denotes the frequency, a and b are empirical coefficients, and S denotes spectral density. Equation 5.2-6 does not account for the imaginary part of the cross-spectrum, which is referred to as the quadrature spectrum. In homogeneous turbulence, the quadrature spectrum vanishes. In structural engineering applications, such as designing the MOB, the quadrature spectrum is assumed to be negligible. No mathematical expressions for the quadrature spectrum appear to be available in the literature. The coefficients a and b in Eq. 5.2.6 are poorly known. Coefficient a varies with wind speed as shown in Figure 5.2.1.3-8, while b varies with both elevation and wind speed, as shown in Figure 5.2.1.3-9.

Inasmuch as the quadrature spectrum is assumed to be negligible, the coherence of longitudinal fluctuations is defined as

$$Coh(r,f) = S^{cr}(r,f)/[S^{1/2}(z_1,f)S^{1/2}(z_2,f)] \qquad (5.2\text{-}7)$$

Measured values of coherence functions are shown in Figure 5.2.1.3-10. Equation 5.2-7 results in nearly unit coherence (i.e., 0.99, say, or more) of wind speed components with very low frequencies (f of the order of 0.05 to 0.01 Hz or less), even if the distance r is fairly large (e.g., of the order of 50 to100 m or more). Recent studies by Mann (1994) and Krenk (1996) tentatively suggest that for very low frequencies the coherence is in fact considerably smaller than unity (e.g., 0.80 to 0.90 or lower) for large r (e.g., r of the order of 50 to 100 m or more).

For the design of a long structure it may be of interest to describe the spectral density of the total along-wind fluctuating force acting on the structure. Existing turbulence models allow such representations to be constructed by integration of pressures over the surfaces of interest (Simiu and Scanlan (1996), pp. 499-502).

For information on spectra and cross-spectra, see also ESDU (1985, 1986).

<u>Time Variation of Wind Direction</u>

For large-scale extra-tropical storms and storms of the downdraft type there are few empirical data and even fewer systematic statistical studies on the time rate of change of the mean wind direction. Representative values of directional wind speed changes are of the order of 1 degree/(20 s) or less. For example, in a hurricane, the direction can change by 180 degrees in a time interval equal to the diameter of the zone of strongest winds (about 30 km, say) divided by the translation velocity of the hurricane (i.e., about 30 km/hr), so the directional change would in this case be 180 degrees/(30 km/30 km/hr) =180 degrees/(3600 s) = 1 degree/(20 s) as indicated earlier.

5.2.1.4 Probabilistic Distributions of Wind Speeds

Asymptotically, that is, for sufficiently large wind speeds, distributions of extreme wind speeds are currently modeled by using "peaks over threshold" models, wherein all mutually uncorrelated extreme wind speeds exceeding a certain threshold are used to estimate the parameters of a generic extreme value distribution. The threshold should not be too low lest the asymptotic assumption underlying the extreme value theory become inapplicable, nor should it be too high lest the sample size become too small, thereby creating large sampling errors. Any individual record may lead to large sampling errors, for which estimation methods are available. However, analyses of large numbers of records expected to have similar probabilistic behavior yield results that, statistically, "wash out" the sampling errors of the individual records and provide credible evidence of the type of distribution that best describes that behavior. According to recent results, reverse Weibull distributions, which have limited upper tail (i.e., for which the probability of exceedance of certain relatively large wind speeds is zero) are a credible model for both extra-tropical and hurricane wind speeds. For details, see Simiu and Scanlan (1996), pp. 612-614, 624-626.

The generalized Pareto distribution (GPD) with tail length parameters c=0, c>0, and c<0 is related to the Type I, Type II and Type III extreme value distributions. (The Type III distribution of the largest values is also known as the reverse Weibull distribution.). The expression of the GPD is

$$G(y) = \mathrm{Prob}[Y \leq y] = 1-\{[1+(cy/A)]^{-1/c}\} \quad a>0, \quad (1+(cy/A))>0 \qquad (5.2\text{-}8)$$

Equation 5.2.8 can be used to represent the conditional cumulative distribution of the excess Y=V-u of the variate V over the threshold u, given V>u for u sufficiently large; c and a are distribution parameters. Computer programs that estimate the parameters of the appropriate distribution corresponding to a given set of extreme wind speed data are accessible in public files available from NIST (National Institute for Standards and Technology) as indicated below.

The mean return period R, in years, of a given wind speed is defined as the inverse of the probability that that wind speed will be exceeded in any one year. Expressions are now presented that allow the estimation from the GPD of the value of the variate corresponding to probability 1 - 1/(λR), where λ is the mean crossing rate of the threshold u per year (i.e., the average number of data points above the threshold u per year), and R is the mean recurrence interval in years. We have:

$$\mathrm{Prob}(Y < y_R) = 1 - 1/(\lambda R) \quad (5.2\text{-}9)$$

$$1 - [1 + cy_R/A]^{-1/c} = 1 - 1/(\lambda R) \quad (5.2\text{-}10)$$

$$y_R = -A[1 - (\lambda R)^c]/c \quad (5.2\text{-}11)$$

$$V_R = y_R + u \quad (5.2\text{-}12)$$

where V_R is the R-year wind speed (e.g., V_{50} =50-year speed) and u is the threshold used in the estimation of c and a. For "epochal" sets consisting of the largest annual wind speeds, $\lambda = 1$. Note that, given u, λ, c, R and V_R, Eqs. 5.2-11 and 5.2-12 yield the parameter A needed for the estimation of V_R. These expressions are used in the computer programs accessible as indicated in the public files available from NIST as indicated below.

Expressions for joint probability distributions of extreme wind speeds and wave heights do not appear to be available. However, methods based on hindcasting may be considered for estimating joint probabilities at specific locations.

There are few extensive sources of wind speed data readily available to the public. Wind speed data recorded in extra-tropical storm regions and simulated hurricane wind speed data for the Gulf and Atlantic coasts are accessible in public files issued by NIST (see below). Some information is available for the Caribbean islands, East Asia and other areas of the world, but most of it appears not to be readily available. NAVFAC DM 2.2 (1981) and Structural Design Criteria, Loads (1988) contain some information on extreme wind speeds in various parts of the world. Both these sources are weak and provide minimal information that may be difficult to extrapolate to ocean conditions.

Instructions for Accessing Data and Computer Programs Issued by NIST

To access data and programs type first:ftp ftp.nist.gov; >user anonymous; enter password >your e-mail address; >cd/pub/bfrl/emil. This places you in the main directory. Datasets and programs are stored in three subdirectories named maxyear, hurricane and directional. Each subdirectory has a *readme* file.

For example, to access the *readme* file for the hurricane directory, type from the main directory: >cd hurricane; >get readme. To get back to the main directory, type cd ../

To access hurricane datasets or programs, from the main directory type: cd hurricane/datasets (to access datasets) or cd hurricane/programs (to access programs). Then type >prompt off;>dir;>mget * (this copies all the data files). Once you are in the subdirectory hurricane/datasets, if you wish to get a specific file, type: get <nist.name> <local name> (example: get file35.dat file35.dat). To finish the session type: >quit

5.2.2 Currents

5.2.2.1 Measurements

Direct measurements of ocean currents have not been made for as long a time or as at many locations as the winds. A brief review of how the measurements have been made is provided here. They fall into two types: Eulerian and Lagrangian. Eulerian flow measurements are time series of velocity past a fixed point collected, for example, by a moored current meter. Lagrangian velocities are time series of the drift rate of a particle as it moves along; such velocities can be collected from satellite-tracked drifters.

Some of the earliest measurements of the surface currents were of the Lagrangian type, made as estimates of ship drift due to surface currents. Later, dedicated surface drifters were used. At first, these were drift bottles, providing launch and recovery locations. The latest drifters are satellite tracked, with drogues set at specific near-surface depths, such as 15 meters and minimal windage, so that they provide relatively accurate (within 5 cm/sec) estimates of near-surface currents. Such drifters are being distributed globally on a routine basis (Niiler et al. (1987), Figure 5.2.2.1 - 1) as they are also a necessary means of providing ground truth sea surface temperatures for calibrating satellites. These Lagrangian drift measurements, historically and at present, are of use to provide maps of surface currents and to indicate the presence of features, such as eddies and coastal jets, embedded in the mean field or localized to a specific region.

Eulerian velocity measurements provide time series of velocity variability at a fixed point, which may then be used to examine the forces on a structure at that point. Instruments, known as current meters that measure currents at a fixed point, were developed early this century by Ekman (1905, 1926) and others. Initially, currents meters used mechanical sensors, such as propellers or rotors to measure speed and vanes to measure flow direction; these early current meters were found to do a poor job of measuring surface currents accurately because they did not properly average the high amplitude, high frequency fluctuations in velocity associated with surface waves and/or the surface mooring to which they were attached. In addition, the oceanographic community had little success in maintaining instrumented surface moorings for periods of one month and longer until the mid-1970s, so early Eulerian surface velocity measurements are rare.

In the mid-1970s current meters were developed that provided accurate near-surface velocity measurements, both of the relatively small mean flows (as small as several cm/sec in the middle of the ocean gyres) and also of the high amplitude orbital velocities of the surface waves. The propeller-based Vector Measuring Current Meter (VMCM) (Weller and Davis (1980)) has been used on surface moorings deployed at various sites around the world on surface moorings and has provided accurate Eulerian time series of surface currents at depths as shallow as 5 meters. Effort to develop current meters with no moving parts continues at present, as they would be less susceptible to biofouling and breakage. Electromagnetic current meters that sense the electric field associated with sea water moving through the earth's surface have become available, but have not yet become widely used due to issues about reliability and accuracy. Acoustic travel time current meters that measure the difference in the time for sound to travel with and against the flow are also being developed, but have also not been established as proven performers near the surface.

Acoustic current meters that sense the Doppler shift of sound reflected off particles in the water have been developed for use on moorings and ships. Range-gated Doppler current meters can provide a profile of the water velocity along the acoustic beam, and the combination of several beams can be used to resolve a vertical profile of vector velocities or to provide a map in a plane of the radial flow. Downward looking acoustic Doppler current profilers (ADCPs) mounted in the hulls of research ships are now run routinely and provide an ongoing source of information about currents in the upper ocean. Prior to installation of ADCPs in ships, some vessels towed an electromagnetic current measuring device, a Geomagnetic Electrokinetograph or GEK, to obtain surface currents while underway.

Satellite altimeters can determine the elevation of the sea surface. This in turn can be used to determine the geostrophic flow. Satellites record a footprint or swath width of 2 to 12 km wide, depending on which satellite and sensor the data come form. Thus, perhaps the best resolution you would get would be to average in time along the flight path to get the signal from a square area, between 2 km x 2 km and 12 km x 12 km. There is also a sea state dependence of the swath width; the SEASAT swath width varies from 2 km in calm seas to over 12 km in seas with up to 20-m significant wave height. Further review of present measurement methods for ocean currents can be found in Dickey et al. (1997).

5.2.2.2 Types of Currents

In general, currents are classified according to their dynamics or the location. Thus, there has been discussion earlier in this document about wind-driven currents, geostrophic currents, tidal currents, and inertial currents and also of currents with geographic tags.

The equations:

$$
\begin{aligned}
\partial u / \partial t - f_c v &= -(1/\rho_w)\partial p / \partial x + (1/\rho_w)\partial \tau_x / \partial z \\
\partial v / \partial t + f_c u &= -(1/\rho_w)\partial p / \partial y + (1/\rho_w)\partial \tau_y / \partial z
\end{aligned}
\qquad (5.2\text{-}13)
$$

are the general equations of motion. Letting the four terms, from left to right, be called the first, second, third, and fourth terms: wind-driven flows are governed by the force balance described by the first (acceleration), second (Coriolis force), and fourth (acceleration due to wind stress) terms; geostrophic flows are governed by a balance of the second (Coriolis) and third (horizontal pressure gradient) terms; time-dependent, geostrophic or quasi-geostrophic flows are governed by a balance of the first (acceleration), second (Coriolis) and third (pressure gradient) terms; inertial currents result from any impulsive forcing, they are the free oscillation that results from a balance of the first (acceleration) and second (Coriolis) terms. Tidal currents arise from additional external forces, the gravitational acceleration of the sun and moon.

Surface currents are a combination of all of the above as are mixed layer currents. Should the mixed layer be shallow compared to the draft of the MOB, there will be vertical shear in the currents experienced by the hull, as the local wind-driven flow does not penetrate below the mixed layer.

5.2.2.3 Current Speeds

In the open ocean, tidal currents are typically 10 cm/sec or less. In coastal waters, especially in restricted passages, tidal flows are much greater. Perhaps the strongest tidal flow in a shipping channel is 8 m/sec in Seymour Narrows in western Canada. The discussion in Section 5.1 provided typical amplitudes of many of the currents.

A common rule of thumb is to consider sub-inertial frequency, wind-driven flows (including surface wave-driven flows) to be approximately 3% of the wind speed. Inertial oscillations can reach 100 cm/sec. Geostrophic flows can, in boundary and frontal zones, range up to 3 m/sec.

5.2.2.4 Directional Variations

Change in flow direction can be anticipated depending upon the type of current. Tidal currents typically revolve around the clock at the tidal period. In coastal waters the flow is biased along shore so the currents execute a more elliptical than circular path. Inertial currents also turn through 360°, but at the inertial period. An eddy moving through a site will cause the current direction to turn through 360°, but the period may be days to months depending on the speed of translation of the eddy. Unstable boundary currents could have similar periodicity in flow direction. Monsoonal currents will reverse every six months. Low period ocean waves, with periods of months, will lead to flow direction changes similar to those associated with large eddies.

Hurricanes and cyclones are strong wind events. The nature of the currents they generate depends on the speed at which they travel over the ocean. Near-inertial oscillations, low-frequency wind drift, and low frequency Rossby waves can be generated. If the hurricane moves slowly, the surface divergence of the mixed layer draws water up from below, forming a ridge. This ridge then later decays by radiating Rossby waves. If the hurricane moves at a speed so that the turning of the wind occurs at an angular rotation rate close to the local inertial rotation rate, then the hurricane will be particularly effective at generating near-inertial oscillations. Thus, the upper ocean currents will change direction according to the sum of (low-frequency wind-driven + near-inertial + Rossby waves).

5.2.2.5 Horizontal Variations

The discussion in 5.1 provided examples of the range of horizontal variations in surface currents to be found, since these will affect the design of the MOB moorings dynamic positioning.

In many open ocean locations, the surface flow will be predominantly the sum of the local wind-driven flow and the geostrophic flow. The wind-driven flow will have the spatial scales of the surface wind field, and the geostrophic flow will have the scales of the density field. Typically, both will have scales of 10 km and larger. A search for small structure in the wind and surface current fields at scale of 10 km and smaller during the Frontal Air-Sea Interaction Experiment (FASINEX, Stage and Weller (1985, 1986)) found variability in the atmosphere associated with small low pressure systems. Variability in the ocean was associated with fronts and small eddies as discussed earlier. At the surface, the boundary of the front may be sharp, with strong gradients of 10s of cm/sec experienced across 100 m horizontally. Once in the frontal flow, the width of the current may be 30 km (Figure 5.1.3-8). Small eddies, as shown in Figure 5.1.3-9 are more localized, with flow changing sign and magnitude by 60 cm/sec in perhaps 5 km. Similar, strong spatial gradients should be anticipated in coastal locations (Figure 5.1.3-7) in association with coastal squirts and jets.

Smaller scales do exist. Nearshore, rip currents, tidal currents and related flows may have scales of meters; it is unlikely the MOB would be in such shallow water. Small-scale flows, such as Langmuir cells (counter-rotating vortices in the mixed layer whose axes are in the horizontal plane, aligned with the wind) are common, but are not strong (Weller and Price (1988)). The Langmuir cells lead to bands of downwind flow with separations of from a few meters to the depth of the mixed layer (50 to 100 m), and the flow in the bands may be 10 to 20 cm/sec downwind relative to the other surface flow. However, the bands are only a few meters in width.

Shipboard ADCPs provide a good way to map surface current variations, and such data is available (Figure 5.2.2.5-1). However, research ships traveling at approximately 10 knots alias the spatial variability associated with tides and inertial oscillation, so care is suggested in working with raw shipboard ADCP data.

5.2.2.6 Time Variations

Typically, near the surface, variability at and near the local inertial period has elevated energy levels. Figure 5.2.2.6-1 shows velocity spectra from select depths from a mooring deployed south of Iceland. The 10-m spectra is in the correct position, the 102 m is offset 2 decades below the 10 m, and the 198 m is offset 4 decades below the 10 m.
The inertial currents in the northern hemisphere would be clockwise rotating, and a broad band around the local inertial period (marked with *f)* shows energetic motion. This is locally-excited wind-driven motion. Note that it is not found at depths much greater than the thermocline, so would not expect to be seen below about 1000 m here. There is also a tidal peak (M_2) that persists with depth. Tidal flows are found principally at diurnal and semi-diurnal periods. The strongest tidal components would have periods between about 12 and 24 hours.

Energy levels in the spectra rise at lower frequencies. Near the surface, flow variability at periods of 5 to 20 days should be expected as the local response to synoptic weather. Lower frequency variability associated with eddies and other quasi-geostrophic flow would also be expected. Velocity spectra typically increase with decreasing frequency, as shown in Figure 5.2.2.6-1, and a rule of thumb is that the spectra include the point with coordinates (1 $(cm/sec)^2$/cph, 1 cph).

5.2.2.7 Current Models

Various models are reviewed in discussing a general strategy for producing time series of velocities in the hindcasting section (Section 5.3.4). Here, it should be pointed out that there are also purpose-built models in the oceanographic research community. For example, there are models of the upper ocean response to hurricanes and there are regional models of the flow in locations such as the Arabian Sea. At present, any use of models to produce ocean surface velocities requires cross-checking the model currents with historical data to establish their validity.

5.2.3 Wind Waves

5.2.3.1 Introduction

The interface between atmosphere and ocean is characterized by a sharp change in the fluid density by a factor of about 800. Any disturbance to the position of this interface is resisted by two restoring forces: gravity and surface tension. The former is the dominant force for all but the shortest waves (< 2 cm in length) and consequently is the only one that need concern us here. These "surface gravity waves" are generated by several different mechanisms:

1) gravitational attraction of heavenly bodies (principally the sun and moon); these are the "tides" or "tidal wave" (not tsunamis)
2) sudden displacements of parts of the sides or bottom of ocean basins due to earthquakes, volcanoes or landslides; these are "seismic sea waves" or "tsunamis"
3) normal and tangential forces produced by the wind flowing over the sea; these are "wind-generated waves" or "wind waves" and are further categorized as "wind-sea" and "swell". Wind-sea are waves generated by a local storm or wind event, whereas swell has traveled outside of its generating area and may have come from a distant part of the water body.

Tidal waves and tsunamis are generally not of much engineering significance in the deep ocean but, intensified by the reduction in water depth at the approach to land, they can produce elevation changes near the coast of tens of meters. These are long waves having periods of minutes (tsunamis) to up to about 12 hours (tides). Wind waves on the ocean have much shorter periods from fractions of a second to about 25 seconds. Long wind waves generated by extensive and violent storms can reach heights of 30 m and more and are the principal agent of inconvenience, danger and damage to offshore operations.

These three types of surface gravity waves differ markedly also in their levels of predictability. Tidal waves, on an otherwise quiescent ocean, are precisely predictable since they respond deterministically to the regular motions of the nearby astronomical bodies. In the real ocean, the level of predictability is reduced by interaction of the tides with far more random currents generated by wind, internal waves and density differences. Tsunamis are no more predictable than the earthquakes, volcanoes and landslides to which they owe their genesis. The theory and practice of predicting wind waves is very well developed (Komen et al. (1994)) and depends essentially on an accurate specification of the future surface wind field over the entire water body (forecasting) or the past wind field (hindcasting). Modern numerical wave prediction methods yield directional spectra of waves at chosen grid points on a numerical grid. While some offshore engineering applications require wave spectra only, most need other specifications such as the maximum crest height above the mean water level, the run of successive large waves ("groupiness"), the joint distribution of heights and periods and the degree of long-crestedness. This last characteristic may be of fundamental importance to the design and safe operation of very large floating structures such as the proposed MOB.

5.2.3.2 Measurements

Introduction

Measurements of waves at sea may be separated into those that report:

1) the time history of elevation at a point or various statistics derived from this

2) elevation and other surface variables such as the slope vector or current velocity vector from which certain statistics of the direction of propagation of wave energy may be deduced

3) images of the surface elevation or slope over an area of large linear dimension compared to the wavelength of the longest waves, from which information may be deduced on energy propagation directions and coherence of lines (or curves) of extrema (crests and troughs in the case of surface elevation)

In terms of spectra, the first type yields simple frequency spectra, in which the distribution of energy into frequency bands may be estimated. The second type yields directional spectra, in which the energy is distributed in both frequency and direction. The third yields wavenumber spectra, in which the energy is distributed over the vector wavenumber plane. The magnitude of the wavenumber is 1/L (cyclic wavenumber = number of wavelengths, L, per meter) or $2\pi/L$ (radian wavenumber) and the vector points in the direction of propagation. A single image of the surface yields a "180° ambiguity" in the propagation directions, since waves may be traveling in either direction normal to their crests. Two images, separated in time by a reasonable fraction of a wave period, resolve this ambiguity, and indeed, a continuous stream of such images yields the complete frequency-wavenumber spectrum.

Methods

The most common methods of measuring waves at sea use moored accelerometer buoys, bottom or platform mounted pressure transducers or wave staffs. The accelerometer buoys have the significant advantage that, moored well away from offshore structures, their measurements are uncontaminated by reflection and diffraction from the structure. On the other hand, these loosely moored buoys do not yield the elevation at a point as they are carried back and forth by the orbital velocity of the waves. Since they are carried in the direction of propagation on the crest and in the opposite direction in the troughs, this has the effect of broadening the crests, sharpening the troughs and increasing the apparent mean elevation so that the Stokes-like nonlinearity is reduced in the measured record and the height of the crests above the reported mean water level is also reduced (Magnusson et al. (1997)). Pressure transducers, fixed to the bottom or to a structure, yield a record of pressure at some depth, from which the surface elevation is deduced using linear theory. This extrapolation process is sensitive (exponentially) to the ratio of depth to wavelength and fails for waves shorter than the depth of the transducer (Bishop and Donelan (1988)). Furthermore, when the waves are nonlinear the bound components are assigned (from the linear dispersion relation) wavenumbers that are incorrect and, hence, so are the extrapolations to the surface. Wave staffs attached to a structure are capable of good fidelity if kept clean and several diameters from major surface intersecting structural members. In recent years the development of laser altimeters for surface elevation has simplified the placement of these staffs and eliminated the need for cleaning and protection from flotsam and approaching boat traffic.

Wave directional measurements are most often derived from buoys that respond to the surface slope or horizontal orbital acceleration as well as the vertical acceleration. The measurement of surface slope from a buoy that pitches and rolls involves certain compromises. Ideally the buoys should track the surface slope accurately, but such sensitivity to slope predisposes the buoy to capsize in rough seas. The usual compromise is to moor it in such a way that its response to slope is restricted. The price to be paid here is that the slope measurements have to be corrected and these corrections are usually accomplished by making the assumption that the waves follow linear theory. More recently, tri-accelerometer buoys have been used for directional measurements. These generally have spherical hulls with a vertically stabilized three orthogonal set of accelerometers. The buoy is assumed to follow the orbital motion of the surface, from which can be deduced the wave directional properties. An equivalent approach, which is often used on offshore platforms, is the combined measurement of wave height or pressure with two components of orbital velocity from a current meter. The pressure-current meter combination is most convenient because it is out of the way of surface traffic and debris, but it suffers from the loss of short wave information, as described above.

The other traditional method of measuring wave direction uses arrays of wave staffs (see, for example, Donelan et al. (1985)). These have been largely applied as a research tool since their operational use is generally too risky and expensive to maintain. This restriction may be, to a large extent, avoided by the use of laser altimeters and the advantage of recovering the nonlinear characteristics of the surface is further incentive to the use of laser altimeters.

Marine scanning radars (Young et al. (1985)) have been, for some time, used on oil platforms in the North Sea. These radars respond to scattering from very short (centimetric) waves and the long waves are imaged through their modulation of the short waves. The modulation transfer function, which allows one to interpret these radar images in terms of surface elevation variations, is still a matter of active research. In the meantime, this approach must be regarded as approximate. Other remote sensing methods that depend on the amplitude modulation of short waves by long suffer from the same shortcoming. Synthetic aperture radars (Alpers et al. (1981)) and grazing angle methods such as FOPAIR (McIntosh et al. (1995)), which respond to the Doppler shift produced by the advection of the (centimetric) radar scatterers on the orbital velocities of the long waves, are in principle capable of good fidelity and broad area coverage. Conditions of beam width and range resolution may restrict their use in establishing extremal local values. On the other hand, they may be the best approach to examining such important properties as groupiness and long-crestedness.

5.2.3.3 Wave Spectra

Wind-sea

Kitaigorodskii's (1962) similarity theory of wind-generated waves provides the underpinning for most modern descriptions of wind-generated wave spectra. In essence, wind-generated spectra can be described by very few parameters. From observations of waves at long fetch, Pierson and Moskowitz (1964) showed that their spectra approached a common form, in which the propagation speed of the waves at the peak of the spectrum exceeded the boundary layer wind speed by about 20%. It is believed that this is the limiting form for spectra approaching full development, i.e., given unlimited space and time the spectra would asymptote to the Pierson-Moskowitz (P-M) form and there would be no further lengthening of the peak period. If the wind should drop, these waves would be overdeveloped and would become swell.

Hasselmann et al. (1973) pointed out that the observed spectra at short fetch have excess energy over the P-M form near the peak. These young waves (i.e., the propagation speed of the peak of the spectrum is less than the wind speed) are being strongly forced by the wind and they absorb energy faster than they can distribute it to even longer waves. Consequently, they become steeper and the spectrum shows some "enhancement" near the peak. The JONSWAP spectrum (Hasselmann et al. (1973)) describes the enhancement over the P-M form but the enhancement factor seemed to be independent of wave age (A_w = C_p/U, i.e. ratio of propagation speed of waves at the spectral peak, C_p, to the wind speed at 10 m height, U). In consequence the JONSWAP spectrum does not relax to the fully developed form of Pierson-Moskowitz. Subsequently, Donelan, Hamilton and Hui (Donelan et al. (1985) (DHH) showed, from observations at short and moderate fetch, that the peak enhancement depends on wave age. They devised a modified-JONSWAP spectral form that incorporates the wave age dependence of the peak enhancement. This spectral form depends only on the peak frequency and wave age and appears to describe actively wind-generated waves accurately. The DHH spectral form is derived from data in the wave age, A_w, range $0.2 < A_w < 1.2$ and is given by:

$$S(f) = ag^2 (2p)^{-4} f^{-4} f_p^{-1} \exp\left[-\left(\frac{f_p}{f}\right)^4\right] g^{\exp\left[-(f-f_p)^2/2m^2 f_p^2\right]} \qquad (5.2\text{-}14)$$

Where:

$$\alpha = 0.006 A_w^{\ -0.55} \qquad (5.2\text{-}15)$$

$$m = 0.08\left(1 + 4A_w^{\ 3}\right) \qquad (5.2\text{-}16)$$

$$g = 1.7 - 6.0 \log A_w \ : \ for\ 0.2 < A_w \le 1 \qquad (5.2\text{-}17$$

$$g = 1.7;\ for\ 1 < A_w < 1.2 \qquad (5.2\text{-}18)$$

and f is the cyclic frequency and f_p its value at the spectral peak. The spectral parameters (a, m, g) are functions of the wave age A_W only.

Figure 5.2.3.3-1 shows the mean spectra from a set of fourteen 17-minute records from the North Sea in the storm of January 28, 1994. The data are taken from a period of 4 hours, in which the wind speed was steady and the significant height remained within ±50 cm of 8 m, while the estimated peak period varied within ±2.5 seconds of 12 seconds. Much of the variability in the estimates is due to spectral sampling variability (see, for example, Donelan and Pierson (1983); or Bishop and Donelan (1988)); the mean period, estimated from the ratio of the zeroth moment to the first moment, shows far less variability (9.7 ± 0.6 seconds). The average wave age was 0.8 and the figure shows the calculated DHH spectrum for this wave age (dashed line) as well as the relaxed fully developed form, $A_W = 1.2$ and the strongly forced case when the peak phase speed is half the wind speed, $A_W = 0.5$. The bandwidth of the spectral estimates is 0.0078 Hz. and the individual 17-minute sets have 16 degrees of freedom per spectral band so that the 90% confidence limits are quite broad (0.6 to 2.0). The mean spectrum shown has 224 degrees of freedom and its 90% confidence limits are ± 17 % of the values shown. It is clear that this spectrum, which is quite young - wave ages on the open ocean are seldom less than 0.5 except in very sharp spatial or temporal wind gradients – is greatly enhanced near the peak compared the P-M fully developed case. Even greater forcing $A_W = 0.5$, such as might occur in rapidly developing storms, will lead to great enhancement with the largest waves in the system being also very steep.

Swell

The DHH spectrum represents wind seas in all stages of the development from very short fetch or sudden storms to full development. On the other hand, when the wind drops quickly or turns sharply the waves can become over-developed ($A_W = C_p/U \cos\theta > 1.2$, where θ is the angle between the wind at 10 m and the propagation direction of the waves at the spectral peak). At this point the largest waves in the system are decoupled from the wind and they begin to lose energy. There is no similarity form for the spectrum of swell because, linked to no external variable, their spectral shapes are developed over the history of their travels following their escape from the storm that generated them. The shorter waves are attenuated more quickly and the long waves propagate faster, so as the swell travels, it becomes more narrow-banded and there is a steady progression to lower peak periods as a train of swell from a given storm arrives with the longest (wavelength and period) waves at the front. The sea state is a mixture of locally generated wind-sea and swell systems from distant storms. In principle, there can be several distinct incoming swells to the area of interest, but, usually, the competition for delivering large dangerous waves rests with the local wind sea and one dominant swell at a time.

5.2.3.4 Directional Spreading

The spreading of wind-sea about some mean (or peak) direction is usually assumed to be symmetrical about the peak and the rate of spreading is least near the peak and broadens towards low and high frequencies from the peak. This pattern was already evident in the JONSWAP experiment (Hasselmann et al. (1973)) and further work (Mitsuyasu et al. (1975); Hasselmann et al. (1980)) served to confirm it. These measurements were made by slope measuring (heave-pitch-roll) buoys, although in Mitsuyasu's case a cluster of buoys ("cloverleaf buoy") was used to measure surface curvatures as well as slopes. The resolution of directional spreading by such buoys is broad and tends to artificially broaden the estimates of spreading. Donelan et al.(1985) obtained frequency-wavenumber spectra from an array of 14 wave staffs in Lake Ontario. The observed directional spreading is modeled by the sech^2 function, the argument of which is parameterized in terms of the ratio of frequency to peak frequency.

$$f(f,q) = \frac{1}{2} S(f) b \sec h^2 b\left(q - \bar{q}(f)\right) \qquad (5.2\text{-}19)$$

where $\bar{q}(f)$ is the mean wave direction and *b*, the spreading parameter, is given by:

$$b = 2.61\left(\frac{f}{f_p}\right)^{1.3};\ for\ 0.56 < \frac{f}{f_p} \leq 0.95 \qquad (5.2\text{-}20)$$

$$b = 2.28\left(\frac{f}{f_p}\right)^{-1.3};\ for\ 0.95 < \frac{f}{f_p} < 1.6 \qquad (5.2\text{-}21)$$

$$b = 1.24;\quad otherwise \qquad (5.2\text{-}22)$$

Donelan et al. (1985) obtained frequency-wavenumber spectra from an array of 14 wave staffs in Lake Ontario and modeled the directional spreading by the sech^2 function, the argument of which is parameterized in terms of the ratio of frequency to peak frequency. These data cover a wide range of wave ages (C_p/U) conditions from strongly fetch-limited to fully developed. No oceanic data on the directional properties of waves have yet been obtained - or, at any rate, published - with the high resolution of the Lake Ontario data. Consequently, these (Lake Ontario derived) spreading functions are applied to oceanic waves under wave age conditions in the expectation that an increase of the longest wavelengths by about half an order of magnitude is unlikely to alter the form of the spectra of a process that covers at least 4 orders of magnitude in wavelengths.

The shape of the $sech^2$ directional spreading is given in Figure 5.2.3.4-1 for the narrowest distribution (just below the peak frequency) and three other frequencies. The half-height width of the distribution near the peak is 41.5 degrees, so that in a wind-sea the most energetic waves, although concentrated near a given direction, cannot be treated as purely two-dimensional, i.e., infinitely long-crested. Swell is generally more directionally focused and the directional width will vary from case to case – the most narrowly focused having originated in very distant storms – but a reasonable working width is probably about 20 degrees.

5.2.3.5 Statistical Properties of Heights, Periods and Groups

A great deal of success in describing the statistical properties of the surface elevation starts from the assumption that these arise through the independent propagation of a spectrum of waves of various wavelengths. If the components are assumed to be statistically independent then the probability density function, p(η) of the surface elevation variations, η would be Gaussian:

$$p(h) = (2pm_o)^{-\frac{1}{2}} \exp\left(-\frac{h^2}{2m_o}\right) \qquad (5.2\text{-}23)$$

where m_r are the moments of the frequency spectrum S (f) of η:

$$m_r = \int_o^\infty f^r S(f)df \qquad (5.2\text{-}24)$$

The zeroth moment

$$m_o = s^2 = \int_o^\infty S(f)df = \overline{h^2} \qquad (5.2\text{-}25)$$

is the variance of the surface elevation.

The Gaussian assumption appears to be quite closely satisfied for large ocean waves and has been applied successfully as early as 1953 (St. Denis and Pierson) to calculate the motion of ships in a confused sea. Generally speaking, the Gaussian assumption is valid in most deep water engineering applications, where one is interested in waves near the spectral peak and not in the smaller scale waves at higher frequencies. Significant nonlinearities appear in the highest waves and in shallow water (peak wavelength > 4 × the water depth). These nonlinear distortions to the wave profile invalidate the assumption of Gaussianity but, inasmuch as they are only of order 0.1 (related to wave steepness), can be treated as a perturbation on the essentially Gaussian nature of the process. Nonlinearities are discussed further in the next section.

Heights

The envelope of a Gaussian process has a Rayleigh distribution and, if the spectrum is sufficiently narrow - in practice this is generally so for a wind-sea or swell acting alone - the crest-to-trough height may be identified with the double amplitude of the envelope. Thus, the probability density function of heights p(H) is (Longuet-Higgins (1952)):

$$p(H) = \frac{H}{4s^{2}} \exp\left(-\frac{H^{2}}{8s^{2}}\right) \qquad (5.2\text{-}26)$$

or in terms of the normalized height, h = H/σ:

$$p(h) = \frac{h}{4} \exp\left(-\frac{h^{2}}{8}\right) \qquad (5.2\text{-}27)$$

The probability of exceedance of a given normalized height h_+, the cumulative probability, is the integral of (5.2.27) from h_+ to ∞:

$$P(h_{+}) = \exp\left(-\frac{h^{2}}{8}\right) \qquad (5.2\text{-}28)$$

These probabilities are graphed in Figure 5.2.3.5-1 and compared to the measured probability of occurrence of various heights during the storm discussed in Section 5.2.3.2.

If one wants to know the average heights (or heights squared − relevant to force calculations) of waves exceeding h_+, these may be computed from (5.2-27) and (5.2-28):

$$\overline{h_+^n} = \frac{\int_{h+}^{\infty} h^n p(h)dh}{P(h_+)} \qquad (5.2\text{-}29)$$

Heights and Periods

Calculation of wave forces on structures requires the joint distribution of periods, T and heights H. The theoretical joint distribution of normalized heights, *h* and periods, τ (Longuet-Higgins (1983)) is given by:

$$p(h,\tau) = \frac{h^2}{\sqrt[4]{2\pi}\,\nu\left[1+\left(1+\nu^2\right)^{-1/2}\right]\tau^2}\exp\left(\frac{-h^2}{8}\left[1+\left(1-\frac{1}{\tau}\right)^2/\nu^2\right]\right) \qquad (5.2\text{-}30)$$

Where:

$$\tau = \frac{\mathrm{T}m_1}{m_o} \qquad (5.2\text{-}31)$$

and the spectral width, ν is:

$$\nu = \left(\frac{m_o m_2}{m_1^2} - 1\right)^{\frac{1}{2}} \qquad (5.2\text{-}32)$$

The measured distribution during the North Sea storm discussed above is shown in Figure 5.2.3.5-2 and compared with the theoretical distribution (Eq. 5.2-30). The measured spectral width, ν was 0.445 for this storm and the same value was used in the theoretical calculations. The agreement is good, and laboratory tests by Doering and Donelan (1992) further support this theoretical form for deep water and shoaling waves until the Ursell number becomes quite large. The estimates of height and period from the data are deduced by the use of the Hilbert transform of the data. The magnitude of the (complex) Hilbert transform is the envelope of the wave heights (or the wave amplitude or half height for a Gaussian sea), while the time rate of change of phase is instantaneous frequency if the spectrum is sufficiently narrow. Thus this method may be used for pure wind sea (as is the case here) or a single pure swell component, but not for mixed seas.

Periods

This theoretical distribution may be used to obtain other useful statistics such as the most probable (mode) period τ_m associated with a given height, h :

$$t_m = 2\left(1+\left(1+32n^2/h^2\right)^{\frac{1}{2}}\right)^{-1} \tag{5.2-33}$$

shown (dashed) in Figure 5.2.3.5-2; or the probability of normalized periods t regardless of h (Figure 5.2.3.5-3):

$$p(t) = \left(nt^2\left[1+\left(1+n^2\right)^{-\frac{1}{2}}\right]\left[1+\left(1-\frac{1}{t}\right)^2 n^{-2}\right]^{\frac{3}{2}}\right)^{-1} \tag{5.2-34}$$

In Figure 5.2.3.5-3 the two curves illustrate the distribution of t for different spectral widths, ν. The solid curve corresponds to the North Sea wind-sea spectrum of Figure 5.2.3.3-1, while the dashed curve is for a much narrower spectrum (ν = 0.2) characteristic of swell. The narrow spectrum yields a more symmetrical distribution and, correspondingly, the mode is closer to the mean period than in the wind-sea case.

Groups

The grouping of high waves is an important consideration in many engineering problems and there have been many attempts to characterize it (e.g. Nolte and Hsu (1973); Ewing (1973); Hamilton et al. (1979); Kimura (1980); Rye (1982)). Generally, one wants to know the joint distribution of lengths of runs of waves exceeding a certain height. Figure 5.2.3.5-4 is a contour plot of this distribution derived, for this report, from application of the Hilbert transform to the North Sea data. The Hilbert envelope has been smoothed using a running average filter over 9.7 seconds, which is the mean period of this data set. The height, as before, has been normalized by the standard deviation, σ, while the run length, *l*, is the number of consecutive waves exceeding a chosen height. The highest waves occur for a fraction of their period and "long" runs of 2 periods or more are concentrated around heights of half the significant height (or 2σ). Run lengths greater than one period mean that two consecutive crests of that height or greater could cross the observation point. The peak of the distribution along a vertical slice at *l* = 1 is at about 2.5σ or 60% of the significant height and there is a significant probability (the contour interval is 12.5%) of two consecutive crests greater than the significant height. The average lengths of runs of a given height or greater is graphed in Figure 5.2.3.5-5. (Note that the axes in this figure are interchanged from those in Figure 5.2.3.5-4). Also shown is the empirical curve of Ewing (1973):

$$\bar{l} = \left(h_+\right) = \sqrt{2}Q_p h_+^{-1} \tag{5.2-35}$$

where Q_p is Goda's (1970) spectral peakedness parameter:

$$Q_p = 2m_o^{-2}\int_o^{\infty} fS^2(f)df \tag{5.2-36}$$

Ewing's average run lengths are somewhat larger. The reasons for this may have to do with different methods of estimating runs or that the data were obtained in very different conditions. This issue will be examined in the quantitative stage of the work for the MOB application.

5.2.3.6 Spatial Coherence

The topic to be addressed here is the reliability of present methods of modeling waves in predicting, from detailed wave measurements at one point, wave characteristics at another point considerably distant from the first. By distant we mean distances of the order of about one to perhaps ten wavelengths, which can be the scale of the MOB.

For the simplest case of waves traveling in only one direction (long-crested waves), we can ask the following questions: Firstly, how closely do the wave histories at a reference point match those at distant points located perpendicular to the direction of wave travel? Secondly, how well can we predict, using either linear or nonlinear wave theories the wave histories at a distance downwave from the data at the reference point?

For directionally-spread seas, the situation is more complex, but the question remains. How accurately can we predict the wave histories at any point located several wavelengths away from the reference point, from data at the reference point?

A workshop (ONR (1997)), sponsored by the MOB program, was recently held in Washington, D.C. to investigate the state of knowledge of wave coherence. Its conclusions were the following:

The current status is that there are almost no extensive quantitative observations on spatial coherence of waves in the ocean, although the instrumentation exists to perform these studies. Until such studies have been performed in ocean conditions, it will not be possible either to confirm that linear models are adequate, or to validate more complicated nonlinear theories. Wave tank measurements are of value, but due to scaling problems, validation is necessary in the ocean itself.

Although wave theories exist that include nonlinear processes in long-crested waves, no well-accepted nonlinear models exist for three-dimensional (short-crested) seas. Development of three-dimensional nonlinear wave theories is needed, along with accumulation and analysis of measured ocean data.

5.2.3.7 Linear and Nonlinear Ocean Wave Kinematics

Accurate knowledge of free surface elevations and wave kinematics is needed for the computations of wave forces on members. Given the size of the MOB, the designer must be able to predict the propagation of waves from one end to the other. The following discussions are about the computation of wave kinematics and wave propagation under survival wave conditions, that is, of extreme steep ocean waves.

Up to now, the majority of design practice has been limited to the unidirectional irregular ocean waves. Hence, we introduce methods commonly used by the offshore industry for the computation of wave kinematics and wave propagation in the context of unidirectional irregular waves. These methods have been widely used in offshore structure design for a long period and are well known. Our attention here focuses on their deficiencies when ocean waves are extremely steep. For accurate prediction of wave kinematics, a newly developed hybrid wave model (HWM) is recommended. Finally, the effects of ocean wave directionality or short crestedness on wave kinematics are explored.

Historically, irregular wave characteristics and propagation are calculated using the fast Fourier transform (FFT) spectral method and linear wave theory. It had been found that the predicted particle velocities near the surface of steep ocean waves are not accurate (Wheeler (1970)). To improve the accuracy, two different methodologies have been adopted. Belonging to the first methodology are a variety of modifications to linear wave theory near the free surface, specially in the region of wave crests above the still water level. The modifications are based on the techniques known as “stretching” and “extrapolation”. In the past three decades, numerous modification methods have been proposed based on either stretching or extrapolation technique, or their combination. The most widely used and well known ones are the Wheeler stretching and linear extrapolation methods (Rodenbusch & Forristall (1986)). In comparison with linear wave theory, the two methods improve the prediction of wave kinematics. Nevertheless, they are not based on sound hydrodynamic principles. It is not surprising to find out:

1) in the case of extreme steep ocean waves, based on the same measured wave elevation the predicted wave kinematics obtained using the two different methods can differ by more than 50%, which is of concern in computing wave loads on offshore structures (Steele et al. (1988));

2) the predictions made by the both modifications show very large discrepancies with respect to corresponding measurements (see Figures 5.2.3.7-1 and 5.2.3.7-2). In general, the Wheeler stretching under-predicts horizontal velocities under wave crests while the linear extrapolation over-predicts them (Zhang et al. (1991, 1996a), Randall et al. (1992)).

The second methodology is to approximate the wave kinematics of a steep wave crest in an irregular wave train by that of a regular (i.e., periodic) wave crest which has the similar wave height and wave period (or wavelength) as the irregular wave crest (Bosma and Vugts (1981)). The irregular waves are represented, wave-by-wave, as regular waves, with the height set equal to the difference in elevation of the irregular wave crest and adjacent troughs, and the period set equal to the time between zero up-crossings in the irregular wave. Because the kinematics of regular (periodic) waves of extreme steepness can be accurately calculated using high-order Stokes theory or high-order nonlinear wave numerical schemes (Dean (1965), Cokelet (1977)), it was thought that the wave kinematics under a steep irregular wave crest can be approximated by that under a similar regular wave crest with fairly good accuracy. However, measured particle velocities have shown that kinematic characteristics of irregular waves are different in two respects from those of regular waves with similar wave periods and heights. The horizontal velocity under an irregular crest is much greater near the surface, decays faster below the surface and is smaller than that of a similar regular wave crest as shown in Figure 5.2.3.7-3 (Kim et al. (1990), Zhang et al. (1992)). Secondly, at the same depth from the still water level, the magnitude of horizontal velocity under a crest is usually smaller than that under an adjacent trough in an irregular wave train, whereas the difference is insignificant in a regular wave train. These differences (especially the first one) in wave kinematics between regular and irregular waves with similar wave periods and heights may have important implications on the computation of wave loads. For example, in the computation of wave impact on the upper deck by a steep wave crest, the use of wave kinematics of a similar regular wave crest may result in a significant underestimate of the wave loads.

Because of the very large size of the MOB, the accurate prediction of the propagation of irregular wave elevation over the length or width of the MOB becomes crucial to the computation of wave phases and in turn the total wave loads on MOB. It is thus necessary to examine the calculation of irregular wave propagation using the FFT spectral method based on linear wave theory. Linear phase velocity of each wave components is employed to predict the propagation of its elevation in time or linear dispersion relation to predict its elevation in space. The resultant irregular wave elevation is the superposition of all wave components. The effects of the interactions among wave components on their phase velocity are simply ignored. This simplification may lead to errors in two respects. In an irregular wave train consisting of many wave components, the short-wavelength (or higher frequency) wave components ride on the long-wavelength (or lower frequency) wave components as if they ride on an unsteady current. The particle velocity of the long-wave components have the Doppler effect on the frequencies of the short-wave components, and therefore the phase velocities of the short-wave components changes along the surface of the long-wave components. Secondly, the average frequency of a wave component may increase due to the nonlinearity of its own amplitude and the amplitudes of other wave components. The nonlinear dispersion relation for wave components in an irregular wave train is given below (from Zhang et al. (1996a)). For simplification, the water is assumed to be deep.

$$f_i^2 = gk_i\left[1 + a_i^2 k_i^2 + 2k_i \sum_{j=1}^{i-1} k_j a_j^2 \frac{f_j}{f_i} + 2k_i \sum_{j=i+1}^{M} k_j a_j^2 \frac{f_i}{f_j}\right] \qquad (5.2\text{-}37)$$

where f, k and a denote the frequency, wavenumber and amplitude of a wave component. The subscript i is an integer $f_i = i * \Delta f$. Δf is the frequency increment used in the spectral analysis. The second term in the brackets of the above equation describes the nonlinear effect of its own amplitude as in a periodic wave train. The third and fourth terms result from the nonlinear amplitude effects of the wave components of frequencies lower and higher, respectively, than f_i. The nonlinear correction in wave frequency for a long-wave components in an irregular wave field is usually less than one or two percent. With the increase in the frequency of wave components, the nonlinear correction increases. For a short-wave component whose frequency is about twice of the spectral peak frequency, the nonlinear correction can be five to six percent of its intrinsic frequency. For accuracy, the nonlinear dispersion relation is recommended in the computation of wave propagation and wave loads on the MOB.

Ocean waves may be viewed as consisting of many free-wave components and bound-wave components of different periods, amplitudes and advancing in different directions. The bound-wave components result from nonlinear interactions among the free-wave components. The wavelength and period of each free-wave component obey the dispersion relation and those of a bound-wave component do not. The amplitudes of bound-wave components are, at most, of second order in wave steepness of the interacting free-wave components, that is, the former are much smaller in amplitude. When ocean waves are not steep, the bound-wave components or nonlinear interactions among the free-wave components are trivial and can be neglected. Under these circumstances, linear wave theory is valid. When ocean waves are steep, the interaction between the free-wave components are significant and can not be neglected. The effects of nonlinear wave interactions on wave characteristics are briefly described below:

1) In the presence of the long-wavelength (long) wave components, the “still” water level with respect to the short-wavelength (short) wave components is actually at the long-wave surface. The change in the “still” water level with respect to short-wave components is extremely important in the computation of wave kinematics and dynamic pressures.

2) The short-wave components riding on the surface of long-wave components are modulated by the long-wave components. They travel faster and are shorter in wavelength and greater in amplitude at the crest and slower, longer and smaller at the trough of the long-wave components. These effects present complication to the decomposition of an irregular wave train into free-wave components.

3) The amplitudes of all free-wave components may have effects on the average frequency of each free-wave component, especially on relatively high-frequency wave components.

4) The most significant bound-wave components result from the sum- and difference-frequency interactions between free-wave components of relatively large amplitudes, which are found to be near the spectral peak in the frequency domain. These resulting bound-wave components are in the second-harmonic frequency band and sub-harmonic frequency bands. Although the amplitudes of these bound-wave components

are much smaller than the free-wave components near the spectral peak, they can be comparable to or even greater than the free-wave components whose frequencies are in the second-harmonic and sub-harmonic frequency bands. When the resultant wave characteristics or the responses of offshore structures are sensitive to the contribution from these frequency bands, the neglect of nonlinear wave-wave interaction may result in qualitative or large quantitative errors (Zhang et al. (1996b), Cao and Zhang (1997)).

Nonlinear wave-to-wave interactions in irregular waves are considered in the HWM at least up to second order in wave steepness (Zhang et al. (1996a)). The HWM divides a spectrum into several frequency bands. For convergence, there are two different approaches: the conventional perturbation method and phase modulation method are used to formulate the wave-wave interactions between the free-wave components with close frequencies (located in the same band) and with significantly different frequencies (located in the different bands), respectively (Zhang et al. (1993)). The uniqueness of the HWM with respect to other nonlinear numerical schemes is that it decouples nonlinear bound-wave components from the free-wave component during the decomposition of a measured wave record. The input to the HWM can be a wave record measured at a fixed point, say the wave elevation. In contrast to the FFT, it decouples the nonlinear bound-wave components from the measurements and then renders the free-wave components. Since the computation of bound-wave components needs the information of the free-wave components, the decomposition is accomplished through an iterative process. In the case that the input is not from a wave record, the free-wave components can be alternatively given by the user. Once the free-wave components are known, the resultant wave characteristics and wave elevation nearby the measurement can be calculated by the superposition the contributions from all free-wave components and bound-wave components. The HWM has been validated through extensive comparisons with laboratory measurements (Spell et al. (1996)). Figures 5.2.3.7-1 and 5.2.3.7-2 shows excellent agreement between the predicted particle velocities using the HWM and related measurement.

Although ocean waves are short-crested and wave directionality has important effects on wave kinematics (Forristall 1981), the studies involving both wave directionality and nonlinearity are very few and limited to statistical approaches. The HWM was recently extended to allow for wave directionality and the directional HWM provides a deterministic study, which considers both wave directionality and nonlinearity (Prislin et al. (1997)).

Preliminary results showed that wave directionality plays an important role in wave kinematics and the comparisons of the predictions of the directional HWM with field and laboratory measurements are satisfactory (Prislin (1996), Zhang et al. (1997)).

5.2.4 Internal Waves, Solitons and Fronts

5.2.4.1 Linear Internal Waves

In mid-latitude oceans, the density of surface waters is typically 1-2% less than that of abyssal water. The vertical density gradient, quantified by the Vaisala or buoyancy frequency $N = \left(\frac{g}{\rho} \frac{\partial \rho}{\partial z} \right)^{1/2}$ (where g is the gravitational acceleration and ρ is potential density), supports the propagation of internal gravity waves. Internal waves can exist at temporal frequencies between the Coriolis frequency $f_c = 2\,\omega \sin \lambda$ (where ω is the earth's angular rotation rate and λ is the local latitude) and the Vaisala frequency (typically 0.3 cph in the deep sea, 5 to15 cph in the upper thermocline). Internal waves propagate in the vertical as well as horizontal, transporting energy and momentum vertically over a significant fraction of the ocean depth.

In the open sea the internal wavefield is best described by a continuous spectrum that is red in both frequency and wave number. Deep-sea internal waves have horizontal scales from 200 km (low frequency, low mode) to 100 m or less. Waves of many lengths and directions are found together, producing a confused “internal sea”. In coastal waters the longest internal waves have 20 to 40 km scales. There is frequently a preferred direction, with waves propagating shoreward, crests parallel to sea-floor contours.

The most energetic horizontal currents are found near the inertial frequency. These near-inertial motions are often caused by local wind events. In the aftermath of intense, transient (hours to days) wind forcing, strong (1 m/s for a hurricane) currents can be excited in the ocean mixed layer. The horizontal scale of these motions is large; tens to hundreds of kilometers. The velocity contrast across the base of the mixed layer approaches the full surface speed.

The greatest vertical displacements of the ocean interior usually occur at semi-diurnal tidal frequencies. Within 500 km of sea-floor topography, tidal internal wave variability is often found to be coherent with astronomic forcing. Tidal forcing plays a central role in the creation of solitons (discussed below). In addition to inertial and tidal motions there is a continuous spectrum of smaller scale waves, as well. The total r.m.s. horizontal velocity associated with this continuum is 0.10 to 0.20 m/s at the sea surface (Garrett and Munk (1975, 1979)). The variation in current across the length of a MOB is much smaller, of order 0.02 m/s.

Internal wave currents are uniform in depth in the ocean mixed-layer. The dominant currents vary only slowly with depth below. Small-scale internal wave shear over the depth of the MOB is not a significant design factor.

Several near-surface manifestations of internal waves are of relevance to MOB design. Internal tides can cause ±10 m variation in the thickness of the mixed layer, perhaps rhythmically altering the temperature of engine cooling water. Inertial currents will tend to displace a MOB in a circular trajectory with a radius of 5 to15 km. If the water at the depth of MOB thrusters is stratified, much of the energy applied by the thrusters can go into generating internal waves rather than propelling the MOB. This phenomenon was first identified by the polar explorer Nansen, who termed it “dead water”.

5.2.4.2 Internal Solitary Waves

Solitary waves represent a specific solution to the non-linear internal wave equations in which the dispersive property of linear waves is balanced by non-linear effects. With this balance maintained, isolated disturbances can propagate with unchanging form over great distances. In contrast to linear waves, both the speed of propagation and the wavelength are functions of disturbance amplitude.

The literature distinguishes among solitary waves (non dispersive by virtue of non-linearity), true solitons (which can propagate through complex background environments or through other solitons with no permanent deformation), and internal bores (which interact strongly with the background). Observations show that “solitary waves” are rarely solitary and are usually found in packets of 2 to 10 distinct crests which embrace both soliton and bore-like characteristics (Figures 5.2.4.2-1,2). Henyey and Hoering (1997) have coined the term “solibore” to describe the related set of phenomena. We will use the terms soliton and solitary wave loosely here.

Solitary wave trains are ubiquitous on continental shelves, worldwide. These “shallow water” solitons can have currents in excess of .8 m s^{-1} (e.g., Holloway (1987); Sandstrom and Elliot (1984); Haury, et al. (1979)). However, the wavelength of these disturbances is typically short, 100 to 300 m. They can exert significant stress on a MOB but would not deform the overall configuration of the platform. In contrast, the bore-like aspect of soliton packets on the continental shelf is of a scale comparable to MOB (Figure 5.2.4.2-3). The passage of an internal bore will tend to rotate and displace a MOB.

Deep water solitons are a significant concern. Peak velocities can be as great as 1.5 m s^{-1}, over scales of order 1 km. The longitudinal horizontal displacements associated with these waves often exceeds 1 km. The <u>difference</u> in displacement can exceed 0.5 km over the 1.5-km length of a MOB. This will severely deform a segmented MOB. Non-linear internal waves are associated with the greatest (kilometer-scale) longitudinal strain rates that a MOB will experience.

5.2.4.3 Fronts

Surface fronts represent discontinuities in the horizontal velocity field. Fronts can be controlled by underlying topography or by local dynamical factors. Fronts are often found in regions with strong lateral density gradients, where there is a tendency for the heavier surface waters to subduct beneath the lighter water. (Figure 5.2.4.3-1). These weak vertical motions are commonly associated with strong lateral currents, which flow parallel to the face of the front. Lateral currents can exceed 0.5 m s^{-1}.

Oceanographers have long been familiar with the large scale fronts that bound the major water masses of the ocean. Recently, it has become apparent that the major (200 km scale) fronts are in fact collections of numerous smaller (20 km scale) fronts which in turn are assemblages of highly contorted kilometer scale features. As measurement techniques improve, energetic small-scale fronts are being discovered at sites where no large-scale frontal activity is present. The space-time climatology of these microfronts is poorly known. They are thought to exist primarily in coastal waters.

The lateral shearing of the currents will tend to rotate the MOB. In coastal fronts, this variability can occur over kilometer scales, resulting in significant velocity differential along the structure. Measurements of open ocean fronts tend to show less shear. However, the number of detailed open ocean frontal studies is limited. Fronts are associated with large (kilometer-scale) lateral shear.

5.2.5 Ice

Although ice may be encountered by the MOB during its life, this item is beyond the scope of this project.

5.2.6 Physical Properties of Seawater

Various seawater physical properties such as temperature, salinity and oxygen content may be important for steel requirements, corrosion and buoyancy calculation. Guidance can be found in API RP2A (1993).

5.2.7 Marine Growth

Marine growth or fouling can be expected to occur in most oceans of the world. Barnacles and softer growths adhere to the underwater surface of ships and other marine structures, increasing the effective size of the member for wave loads and changing the hydrodynamic forces, mass and weight of the structure. These growths can be several inches in thickness, but can be kept to a minimum by regular inspection and cleaning.

We will first comment on the static effects of fouling. There is an increase in thickness that results in an increase in effective size of submerged members. This will change all the hydrodynamic forces on these members, but for structural members of dimensions 50 to 100 ft, the effect of a few inches of growth is small. There is also an effect on the weight of the structure, but, partly because the density of fouling is not greatly different from that of water, and partly because the total volume of the fouling is small relative to the displacement of the clean member, this effect is small.

If fouling is allowed to build up, its principal effect would be to increase the drag forces. The increase in drag coefficient is quite important and occurs with quite small amounts or growth. The MOB would then require extra power and hence fuel to drive it at a given speed during transit, or could slow it down, if it is being driven at close to full power. Similarly, there would be a significant increase in thruster power required for station-keeping by the dynamic positioning system, or in the mooring line forces, if the MOB is moored. Increased drag would also have some effect on its dynamic response to waves. However, since drag forces are rather insignificant relative to other wave-related hydrodynamic forces, their effect on the wave-frequency response of the MOB can be expected to quite small.

5.3 Sources of Environmental Data

Information about the surface winds and about wind-generated oceanic phenomena (currents, waves) is available from a number of sources.

Climatologies provide large-scale coverage, low resolution (often monthly means on 2° latitude/longitude grids) of the surface meteorological fields. These are most often derived from the weather reports submitted by ships. The best of the readily available climatologies is the COADS or Comprehensive Ocean Atmosphere Data Set prepared by da Silva et al. (1994). This provides quick access to climatological monthly mean wind speeds, air temperature, sea surface temperature, and other surface meteorological observables; a CD-ROM, *Atlas of Surface Marine Data 1994*, is available from NODC (National Ocean Data Center, see http://www.nodc.noaa.gov/NODC-cdrom.html). The U. S. Navy also has produced surface marine atlases. In particular, there is the eight volume set of the U.S. Navy Marine Climatic Atlas of the World based on data from 1854 to 1978. For example, NAVAIR 50-1C-532 (1979) provides for the South Pacific monthly maps and histogram distributions at select location of surface winds (speed and direction), air temperature, sea surface temperature, humidity, precipitation, visibility, cloud cover, ceiling and visibility, sea level pressure, wave period, direction, and heights, and cyclone movement, wind speeds and directions. These surface marine atlases are also available as a CD-ROM from NAVOCEANO. Pilot charts provide an alternate climatology of surface winds and currents based on ship reports.

Products with global coverage are available from various forecast and analysis centers. The meteorological centers typically provide analyzed fields four times per day (0, 6, 12, 18 UTC) with approximately 1° to 2° latitude/longitude resolution. The European Centre for Medium Range Weather Forecasts (ECMWF), the U. S. National Center for Environmental Predictions (NCEP), and the U. S. Navy Fleet Numerical Meteorology and Oceanography Center (FNMOC) are examples of such centers that provide analyzed fields (i.e., fields of the present conditions in which in-situ observations from ships, buoys, and other sources have been assimilated) as well as forecasts (predictions of future conditions out 36 and 72 hours and further into the future). ECMWF has led the field in this work, and routinely archive the analyzed wave information from their global model. Wavsat (UK) and Oceanweather (US) are two private organizations that develop wave data and statistics from satellite or buoy data. Over the years, the numerical models used by the weather centers have improved, and periodically the centers have reanalyzed past data and produced retrospective, reanalyzed fields. These fields are a valuable resource, as they provide both wind time series at fixed points and information about spatial variability of the wind. A caution when using such fields, though, is that they do not capture small spatial scale and high frequency variability. A small, intense storm or a squall, for example, would not be captured well in such data.

Global Wave Statistics (Hogben et al. (1986)) is a global set of wave data with a sufficiently long observation history to give reliable global wave and wind statistics. This data is from ship observations and goes back to 1854, calibrated against measured buoy data. This data has been fitted to closed form statistical descriptors, so that joint probabilities of significant wave height and zero-crossing period can be estimated.

Satellite sensors now also provide information about surface winds. Active microwave satellite sensors referred to as scatterometers (such as NSCAT, ERS-1) provide vector winds with good spatial resolution (2 km) over a swath of approximately 25 km wide, but it requires approximately 10 days for the ocean's surface to be covered totally by swaths so global fields are composites in space and time. However, because of the good spatial resolution within the swath, these satellite winds are a very valuable resource for examining kilometer scale variability in the surface winds. Local, orographic wind features, such as the strong winds that are seen in the Gulf of Tehuantepec west of gaps in the mountains in central America and flow around islands as well as hurricane and storm winds are well-resolved by the scatterometer. Access to scatterometer data would be through NASA and ESA (European Space Agency). Passive microwave satellite sensors (such as the SSM/I, Special Satellite Microwave Imager) provide global coverage of wind speed. Access to SSM/I data is through NASA.

Radar altimeters on satellites provide information on the significant heights of waves and wind speeds along the satellite track. The GOES satellite, Topex-Poseiden and the ERS satellites have been providing a wealth of data on significant height and wind speed. Unfortunately, there is no information on periods or directions, and the data are limited to a line directly below the satellite. Comments on the resolution of these instruments are given in Section 5.2.2.1. Synthetic aperture radars, on the other hand, cover a wider swath and also yield information on the wavelengths (hence periods) and directions. Such radars have been in use on the ERS satellites for several years and a reasonable climatology is being acquired.

Data servers now provide easy access to some of the climatologies and model products for surface winds. For example, the server at Lamont Doherty Earth Observatory (LDEO), with a web address of http://ingrid.lego.columbia.edu/SOURCES/?help+surface climatologies shows monthly global climatologies of winds, including the COADS product; ECMWF global analyses; NCEP global reanalyses. The Lamont server also provides access to Atlantic shipdrift data. The Surface Currents Data Base Management System SCDBMS) at the Naval Oceanographic Office (NAVOCEANO) provides access the Navy data base of surface current observations. A web server at the NOAA Atlantic Oceanographic and Meteorological Laboratory (AOML) (http://www.aoml.noaa.gov/phod/dac/dacdata.html) provides access to surface drifter data from around the world.

The National Ocean Data Center (NODC, http://www.nodc.noaa.gov/) has CD-ROM compilations of the 1994 COADS data, surface currents from ship drift (including some

GEK observations, but mainly ship drift), global wind and wave data from the Geosat satellite, wind and wave data from the National Data Buoy Center (NDBC) coastal buoys located in coastal U. S. waters, moored ADCP data from the eastern shelf of Florida. The NODC web page also shows links to the World Data Center A, Oceanography and other oceanographic sites.

The Japanese Ocean Data Center (JODC) as well as other national and international data archives also maintain holdings of surface current observations.

The Jet Propulsion Laboratory physical oceanographic data server, JPL PO.DAAC (http://podaac-www.jpl.nasa.gov/) provides a web-based site for access to surface wind products from NSCAT and sea surface elevation and wind products from the altimetric satellite, TOPEX/POSEIDON. The sea surface elevation products provide ocean surface topography, useful for locating currents and eddies, significant wave heights, and wind speeds.

The National Climatic Data Center, NCDC, (a branch of the National Weather Service and National Oceanic and Atmospheric Administration, NOAA) has a long history of providing climatological services with data that goes back 150 years. It has an international role as World Data Center-A for Meteorology and archives satellite, climate and radar data for locations around the world. Data is available on CDROM.

In regions close to shore, the use of radar to map surface currents inferred from the radar backscatter is being explored, but these radars, sometimes referred to as CODAR, are not widely used and cannot be considered as a routine source of data.

Data sources are discussed in more detail below in relation to specific phenomena.

5.3.1 Storms and Wind-Generated Phenomena

5.3.1.1 Typhoons and Hurricanes

Excellent maps of surface winds associated with hurricanes have now become available from satellite scatterometers. Figures 5.3.1.1-1 and 5.3.1.1-2 provide examples of wind fields of two hurricanes. Figure 5.3.1.1-1 shows a closeup of a hurricane over Japan in September 1997.

A second tropical storm was at the same time found east of Japan, and is shown in a series of composites developed from NSCAT overpasses (Figure 5.3.1.1-2). These figures are illustrative of the ability of the scatterometer sensors to capture the wind fields. In a swath spatial resolution approaches 2 km. The satellite carrying NSCAT failed, but a European scatterometer (ERS-1) continues to fly. Past data from NSCAT and past and present data from ERS-1 presents a source of surface winds that can define the spatial scales and translation rates of hurricanes.

Analysis fields from numerical models provide an assimilation of all available surface data consistent with the dynamics built into the model (some models are now assimilating ERS-1 winds as well) and provide another source of wind fields. Near the coast, hurricanes may pass over existing weather buoys (Figure 5.1.2.2-2, for example). Time series would be available, then, from these sites through NOAA National Data Buoy Office (NDBO).

There are few sources of ocean current data taken from under a hurricane. A limited number of aircraft flights have been made in which aircraft expendable shear probes have been dropped in the water; these give the vertical shear of the ocean velocity as they fall through the water column. A surface mooring near Bermuda recently measured surface winds and currents from the vicinity of a hurricane, but the data has not yet become available.

The National Weather service on Guam has tabulated many years of typhoon characteristics in the general region of Guam. A major advantage of this data is the large number of storm events (372) recorded, providing a rich body of information to examine the interrelationships between hurricane parameters. Similar data from the Gulf of Mexico and Atlantic Ocean are available. The data are much more sparse, and limited to more severe storms.

5.3.1.2 Non-tropical Storms

As above, satellite and numerical weather prediction models can provide data fields for extra-tropical storms. There are also a number of coastal and open ocean weather and research buoys that can provide time series at fixed locations. The WHOI surface moorings, for example can provide coincident records of the surface winds and the upper ocean currents. These were listed in Table 5.2.1.1-1 (Locations and Durations of WHOI Surface Mooring Deployments, which provide wind and upper ocean velocity records). Buoy networks are maintained by several countries, e.g., the U.S. National Data Buoy Center of NOAA, the Canadian Marine Environmental Data Centre, France, Britain and Japan, inter alia, maintain active environmental buoy systems. Measurements are routinely made from oil platforms on the continental shelves of such countries as the U.S., Canada, Norway and Britain.

5.3.1.3 Distribution over Possible Transit Routes

Storm track probabilities are available from atlases and climatological data bases as discussed earlier. Analysis of numerical weather prediction fields, such as the NCEP or ECMWF reanalyses that cover periods of many years, can provide a data base from which to extract statistics and specific information about storms along transit routes. A caution in using these is that they may under-represent the severity of these storms or, for small storms, not portray them at all.

Surface currents would come from climatological sources, numerical ocean models, and surface drifter data.

5.3.1.4 Conditions at Deployment Locations

Wind and current conditions at deployment sites could be developed from atlases, historical data bases, and numerical weather and ocean models as discussed above. Scatterometer satellites are also useful.

5.3.1.5 Suggestions for Data for Wind, Waves and Current

Data descriptive of present conditions as well as of the past should be sought from a numerical weather center (NCEP, ECMWF or FNMOC) and from satellites (scatterometer, altimeter, AVHRR, color scanner). These will provide surface winds and indications of the strength of any mesoscale ocean variability. Ocean currents should be sought from historical data. The local winds should be used to drive an ocean mixed layer model and produce local wind-driven currents that should be added to the historical, mean currents.

5.3.2 Non-wind-generated Currents

5.3.2.1 Distribution over Possible MOB Transit Routes

In addition to the numerical models of the ocean currents (see Section 5.3.4 - Hindcasting), data from other sources will be of use in looking at the currents to be encountered along a transit route of the MOB.

Strong mesoscale ocean variability, such as fronts, eddies, and jets, often has a signature that can be remotely sensed. Because the flow is strong and rapid compared to mixing time scales, temperature acts as a passive tracer. So does phytoplankton content. Thus satellite infrared (AVHRR, Advanced Very High Resolution Radiometer) and color scanner imagery can be used to locate such features. Figure 5.3.2.1-1 shows an example of a reconnaissance for ocean fronts carried out in preparation for FASINEX using AVHRR thermal imagery. Similarly, satellite color scanner imagery provides a good means of visualizing coastal upwelling, currents, squirts, and jets because the nutrient-rich water has a biological tracer, phytoplankton, detectable by the satellite. Satellite altimeters can detect the anomalies in sea-level associated with geostrophic features, including eddies. The sea level anomalies can then, under some assumptions, be converted to flow speeds. As they can see through clouds, they provide a valuable means to examine current variability. Off Oman, for example, altimetry did a good job of visualizing the large eddies that drifted down on the region from Socotra.

Surface drifters (Figure 5.2.2.1-1) provide another means to obtain surface velocity estimates. Combined with information about the local wind and satellite altimeter data they could be used to identify the component of the surface flow that is the local wind-driven flow and the component that is the geostrophic flow.

5.3.2.2 Conditions at MOB Deployment Locations

The numerical model, satellite, and drifter data sets discussed above can provide general, quantitative information about the currents. The unique challenge of the non-wind driven flow in terms of developing a design may come from the lack of predictability of the eddies, jets, and fronts. Large-scale features in the non-wind-generated flow fields can be determined from climatology and/or ocean general circulation models initialized with climatological temperature and salinity data; such features are the basin-scale gyres and the boundary currents. The details of where eddies, fronts, and jets are generated and propagate are, however, not well-modeled. Thus, a search of historical surface velocity data as well as of satellite sea surface temperature (AVHRR), altimetry and other data that would shed light on the non-wind-generated mesoscale and smaller-scale variability is suggested in preparation for developing surface current information for use in designing a MOB for a specific site.

5.3.3 Internal Waves and Fronts

5.3.3.1 Geographic Distribution

<u>Solitary Waves</u>

The tide is the energy source for large-scale internal solitary waves. When surface tidal currents impinge on underwater topography, internal tides, as well as less organized internal motions, can be generated. Solitary waves can result from an instability of the larger scale internal tide (Figure 5.3.3.1-1) or through the evolution of a more-or-less random mass of tidally mixed water as it spreads laterally from the mixing site (Figure 5.3.3.1-2).

Deep sea solitary waves have been observed in the Bay of Biscay (New and Pingree, 1990), the Andaman Sea (Osborne and Burch, 1980), the Sulu Sea (Apel et al. (1985)), the South China Sea (Bole et al. (1994)), and the Western Equatorial Pacific (Pinkel et al., (1997)), (Figure 5.3.3.1-3). In spite of their size, solitons are easy to miss in conventional oceanographic data. They have little surface manifestation (a 30-cm bulge in sea surface elevation) and pass through any given site in 30 minutes, faster than the sampling rate of many moored oceanographic sensors. We anticipate that there are numerous other locations where these waves are present.

It is of value to consider sites of <u>suspected</u> soliton activity. As a surrogate for an exact soliton census, we can present a map of the energy loss rate from the surface tide to internal motions (Figure 5.3.3.1-4, Egbert (1997)). This map is based on measurements of tidal sea surface elevation, obtained by the TOPEX–POSEIDON radar altimeter. The dissipation sites tend to be related to the topography of the sea floor. As a second surrogate, Morozov (1995),(Figure 5.3.3.1-5) presents global maps of the growth rate of the internal tide. These are based on the known topography of the sea floor and a global model of predicted tidal currents. The degree of correspondence between the model results and the locations of observed solitons (Figure 5.3.3.1-3) is striking. Table 5.3.3.1-1 presents a list of sites of significant conversion surface to internal tidal energy, as computed by Morozov. The likelihood of deep sea solitons is greater within 300 km of the stronger generation sites.

Table Energy fluxes and internal wave vertical displacements for major underwater ridges

Ridge	Energy flux E_f per 1 m of the ridge length (J m s^{-1})	Total energy flux for the ridge length 10^{10} (J s^{-1})	Vertical displacements ζ (m)	A (m)
Atlantic Ocean				
Reykjanes	1789	0.3	30	40
N.Atl. 50°N–35°N	2937	1.0	30	
N.Atl. 35°N–15°N	6853	1.9	40	36
		near Meteor banks	80	70
N.Atl. 15°N–equator	14079	3.8	60	
S.Atl. equator 25°S	35124	9.5	90	
S.Atl. 25°S–35°S	18835	2.0	67	
S.Atl. 35°S–55°S	4897	1.1	36	
North Scotia ridge & South Sandwich islands	5768	0.6	45	
Walvis ridge	11387	1.9	54	
African–Antarctic	224	0.05	10	
Trindade, Martin Vaz isles	10514	0.2	52	50
Indian Ocean				
West Indian 30°S–50°S	2623	0.08	28	
Maldives	7460	1.2	46	40
Arabian–Indian ridge	6278	2.1	43	
Mascaren ridge	16507	3.6	60	80
Madagascar ridge	1837	0.2	21	
East Indian 5°N–15°S (Ninetyeast ridge)	7352	1.6	44	
East Indian 15°S–35°S	7528	1.6	44	
West Australian ridge	1438	1.9	20	
Central Indian ridge	3362	0.7	29	
Kerguelen ridge	1874	0.3	20	
Macquarie ridge	2988	0.3	32	15
Australian Antarctic	282	0.1	15	
Pacific Ocean				
Emperor	5343	0.9	36	25
Hawaii	2830	0.8	25	
Mariana islands	8982	1.4	45	
Lord Howe rise	4071	0.6	31	
South Pacific 180–140°W	183	0.05	10	
East Pacific 30°S–55°S	2118	0.7	25	
East Pacific equat.–30°S	2857	0.9	29	
Sala y Gomez, Nasca	5838	1.6	41	
Mendocino	4741	0.4	28	20
Marshall, Gilbert islands	9713	2.7	48	
Line, Tuamotu	4236	1.0	33	
Marcus–Nekker	637	0.2	12	
Kusu–Palau	17699	4.1	68	60
Kolwill–Lau	1518	0.3	20	

From Morozov, 1995

Table 5.3.3.1-1. Internal Wave Displacements for Major Underwater Ridges

In addition to the spatial geography of deep sea solitons, there is a regular temporal variability that can be exploited in planning MOB operations. Deep sea soliton generation appears to be very sensitive to the level of tidal forcing. In contrast to shelf solitons, deep sea solitons are formed predominately near the spring tides. For two or three days surrounding the forcing maximum, groups of solitons will be formed at 12.4-hour intervals. In recent Western Pacific observations (Pinkel et al. (1997)), only the strongest of the spring tides was associated with solitary wave formation (Feng et al. (1998)). With proper planning, regions of suspected soliton activity can be traversed during the neap tides, minimizing chances for a soliton encounter.

Fronts

Intermediate and fine scale (10 km) fronts are ubiquitous in the coastal oceans of the world. Offshore, one can expect to find these features associated with the major large-scale fronts, which bound the water masses of the global ocean. An idealized sketch of the frontal structure of the Pacific Ocean is presented in Figure 5.3.3.1-6 (Roden (1975)). The position of even these major fronts is highly variable. Given present day satellite capability, one can expect real-time images of frontal structure to be available on board MOB, in support of routing and station-keeping activities (provided that the fronts have a strong surface temperature signal.)

5.3.3.2 Conditions at Deployment Sites

With respect to the designated MOB deployment sites, the Andaman, South China, and Sulu Seas of S.E. Asia are known deep water soliton sites. Although reports are sketchy, shallow water solitons are thought to be ubiquitous in coastal Chinese waters. While there have been no reports of solitons, frontal activity is known to be active in the Sea of Japan and adjacent waters, where southward flowing waters from high latitude meet the northward flow of the Kurushio Current.

In the North Atlantic, both frontal and soliton activity are prevalent in the shallow waters stretching from Greenland and Iceland through to the United Kingdom and the Irish Shelf. Deep water solitons have been sighted in the central Bay of Biscay, but not further north. They represent a potential issue in MOB transit, but not deployment.

5.3.3.3 Sources of Data

For shallow water solitons, useful references to observations include Halpern (1971); Haury et al. (1979); Sandstrom and Elliot (1984); and Holloway (1987).

Deep water soliton data are presented in Osborne and Burch (1980) (Andaman Sea); Apel et al. (1985), Liu et al. (1985) (Sulu Sea); New and Pingree (1990) (Bay of Biscay); Bole et al. (1994) (South China Sea); and Pinkel et al. (1997) (Western Equatorial Pacific).

A useful review of both data and theory is presented by Ostrovsky and Stepanyants (1989).

5.3.4 Hindcasting of Waves and Currents

5.3.4.1 Hindcasting of Waves

Wave hindcasting is the calculation of wave conditions in the past from a history of the wind over the area of interest. The wind history may come from measured winds or, more commonly, the analyzed (after the fact) products of meteorological numerical models. Wave forecasting is the prediction of waves in the future from weather forecasts of low level marine winds. Wave prediction (hindcasting or forecasting) is of fundamental importance to the offshore engineer because measured wave data sufficient to meet the needs of engineering design is seldom available at the chosen location. On the other hand, meteorological records for many areas of the globe have been systematically archived for nearly 100 years. Unfortunately, the historical data base is not nearly so complete for marine winds and the bulk of the measured marine wind data prior to 1960 was acquired by “ships of opportunity”. The quality of such ship data has often been called into question on three counts:

1) the anemometer is often placed in an area of significant flow distortion especially when the ship is not traveling directly against the wind;

2) ship routes are predominantly in the northern hemisphere, leaving the much larger marine expanses of the south with relatively little data;

3) ships attempt to avoid heavy weather thus introducing a bias towards lower winds.

In recent years (the last two or three decades) numerical general circulation models of the atmosphere have become very reliable and, with the assimilation of high quality *in situ* data from coastal buoys, such as those of the U.S. National Data Buoy Center, NOAA, and satellite sensors (scatterometers and altimeters) that cover the entire marine environment, can be used as a source of analyzed winds anywhere on the globe. Of course, for the marine environment this still is not a long enough record to establish reliable extremal statistics for design conditions (e.g., 100-year winds and waves). It may be that the archives of reliable marine winds could be extended by using the historical terrestrial coastal meteorological records to deduce the historical marine conditions based on the relationship between marine and terrestrial coastal conditions in the more recent past, when the marine analysis has been verified with *in situ* and satellite data. Such a process could be used to produce a long term data base for any chosen area.

Another approach to getting suitable wave statistics in a given area is to measure the local conditions using either *in situ* or remote methods. There are two drawbacks to this approach:

1) one can seldom wait long enough to acquire a sufficient data base from which to compute extremal statistics;

2) the measurements are either a local sample *(in situ*), or an infrequent sweep over a limited area (satellite), both of which have significant sampling variability.

The waves integrate the forcing from the wind so that a correct wave prediction model will yield a smooth wave field in a steady and homogeneous wind field. By contrast, point measurements in space or time will have significant variability, being only samples drawn from the underlying population (see, for example, Donelan and Pierson (1983)). Wave prediction models are now at the stage of development that the errors in predicting wave height and mean period (the jury is still out on the question of directional aspects) are largely attributable to the quality of the analyzed wind (Komen et al. (1994)). This suggests that the best route to long term wave statistics is to acquire good historical wind data and run it through an appropriate wave prediction model.

Local Steady State Wave Hindcasting Model

When the wave conditions of interest are generated locally (i.e. there is no significant swell) by winds that can be regarded as steady and homogeneous, the application of simple fetch- and duration-limited formulae to deduce the overall statistics of the wave field may be appropriate. Many of these wave prediction methods are in use today. Perhaps the best known is the SMB method (Sverdrup and Munk (1947); Bretschneider (1958)), but this has now been replaced by the JONSWAP formulae (Hasselmann et al. (1973)) in the Shore Protection Manual (1984). These methods assume that the wave direction is coincident with the wind direction but Donelan et al. (1985) showed that this is not so in general. Instead the fetch distribution about the wind direction determines the actual approach direction of the most energetic waves. Fetch is the over-water distance over which the wind blows to the point of interest and can be: the distance to the upwind shore (fetch-limited by the basin's geometry); the upwind distance to the far edge of the storm (fetch-limited by the extent of the storm); or the smallest of the above fetches and the fetch corresponding to the duration limit. The reason for off-wind propagation of the energetic waves is easily explained. Consider the fetch-limited case where the wind is blowing at an angle to a straight coast such that to the right of the wind the fetch increases with increasing off-wind angle, whereas to the left it decreases. If the increase in fetch to the right more than compensates for the reduced wind component ($U\cos\theta$), the waves in the off-wind direction to the right of the wind will be larger than those propagating directly in line with the wind. The mean wave direction will thus be biased towards the direction of increasing fetch away from the wind direction.

The predictive rules have been given by Donelan (1980). A relation for the dominant wave energy direction (ψ) versus wind direction (φ) in deep water can be determined for any given location with fetch distribution F_ψ by maximizing the expression:

$$\cos\theta \, F_\psi^{0.426} \tag{5.3-1}$$

where $\theta = \psi - \varphi$ is the angle between wave and wind direction.

For any wind direction (φ), vary the hypothetical wave direction (ψ) until the above expression (5.3-1), with fetch F_ψ being the fetch in the wave direction, is a maximum. This direction (ψ_{max}) is the direction from which the waves will come at steady state. Hereafter the subscript (“max”) is dropped.

Once θ and F_ψ have been determined, the fetch-limited expressions for significant height, H_s and peak period T_p can be applied:

$$H_s = 0.00366\, g^{-0.62} (U_{10} \cos\theta)^{1.24}\ F_\psi^{0.38} \tag{5.3-2}$$

$$T_p = 0.054\, g^{-0.77} (U_{10} \cos\theta)^{0.54}\ F_\psi^{0.23} \tag{5.3-3}$$

where U_{10} is the wind speed at 10 m height.

The possibility that particular cases could be duration-limited or fully developed should be checked. This is done by comparing the fetch with the fetch corresponding to the duration of the wind event, t_d and with the fetch corresponding to full development. These are respectively:

$$F_\psi = 0.012\, t_d^{1.3}\, g^{0.3}\, (U_{10} \cos \theta)^{0.7} \qquad (5.3\text{-}4a)$$

$$F_\psi = 9.47x10^4\, g^{-1}\, (U_{10} \cos \theta)^2 \qquad (5.3\text{-}4b)$$

The appropriate fetch to use in each direction in maximizing Eq. (5.3-1) is the smallest of the geometric fetch and the duration (5.3-4a) and full development (5.3-4b) limiting values.

This procedure is easily computerized and yields accurate significant height, peak period and direction of travel of the energetic waves near the peak. If spectral information is required, the generalized spectral form of Sections 5.2.3.2 and 5.2.3.3 may be applied.

A Time Dependent Spectral Wave Hindcasting Model

In order to handle general situations where neither stationarity nor homogeneity may be assumed, finite difference time stepping models have been developed. The first of these models, which are all based on a "spectral transport equation", is due to Gelci et al. (1957). Since that time various levels of complexity (or completeness in the view of the model developers) have been introduced. The most widely tested and used model today is the WAM model (Komen et al. (1994)).

It is based on the energy balance equation (e.g., Komen et al. (1994)) in terms of the frequency direction spectrum, $F(\omega,\theta)$:

$$\{\partial/\partial t + C_g\, \partial/\partial x\}\, F(\omega,\theta,x,t) = S_i + S_n + S_d \qquad (5.3\text{-}5)$$

where C_g is the group velocity and the "source functions" $S(F;\omega,\theta)$ are respectively: wind input, nonlinear wave-wave interactions and dissipation.

The model is driven by U_{10} applied on a grid of typical spacing of 1° latitude and longitude for the full global model, but the resolution in space is limited only by computer resources. Smaller regional models have typically much finer resolution. This sort of model is referred to as "coupled" because the spectral components are linked by nonlinear wave-wave interactions. The coupling allows the model to handle wind-sea and swell in much the same way with energy flowing between different components through the nonlinear interactions. Validation tests of this sort of model by many different groups show good skill for significant height and mean period.

Relatively little testing of the directional aspects of these models has been reported to date - perhaps because the measurements of wave directional spectra are difficult and not as common as the frequency spectra (see Section 5.2.3.1). Perhaps the most serious limitation of this sort of wave prediction scheme is the directional resolution. Typically, the spectra are divided into frequency and direction bins. The frequency bins are usually fine enough and generally logarithmically spaced to equalize the energy in the bins. The direction bins are typically 12°or 15°. This level of directional resolution is clearly inadequate for the MOB specification.

5.3.4.2 Hindcasting Currents

Given the scarcity of actual observations of currents in the upper ocean, an alternate method of generating realistic currents may be required for use in engineering studies in support of the design of the MOB. This would also give the ability to run simulations using observed or hypothetical winds and currents generated in this manner. Such approaches are discussed here.

In the open ocean, away from the eddy and geographically localized variability of the boundary currents, the upper ocean flow field is to first order a linear combination of the large scale, density driven (or thermohaline), geostrophic flow and the locally-forced, wind-driven flow field. Here, an approach would be to estimate the local wind-driven flow and the geostrophic flow separately.

That wind-driven flow is located within the upper, well mixed layer (see Figure 5.1.1-7, which shows an isothermal layer of approximately 60-m deep). The depths of the mixed layer varies under the combined influence of the local wind and the local air-sea heat flux. If time series of both wind and either of the air-sea heat flux or the variables needed to compute the air-sea heat flux (air temperature sea surface temperature, incoming shortwave radiation, incoming longwave radiation, barometric pressure, and humidity) are available (from a moored buoy or from numerical weather prediction models such as run by NCEP and ECMWF) a mixed layer model is run that predicts the temporal evolution of the mixed layer, including its depth and the currents to be found there. The model of Price et al. (1986) is computationally efficient, available, and freely distributed as Fortran code that has been run on computers from mainframes to desktops (contact Dr. Jim Price, Woods Hole Oceanographic Institution). The resulting velocities are available at the time step and vertical resolution used in the model run (typically 1m vertical resolution and every 15 minutes). Each time step, the air-sea heat flux is applied at the surface, the upper ocean mixed to satisfy static stability, and then the momentum from the wind is applied at the surface and the layer again mixed to satisfy shear stability criteria. In low wind, sunny days, the heat fluxes leads to a shallow mixed layer and wind-driven momentum is trapped near the surface, where moderate wind-driven currents are seen. Under strong heat loss, a deep mixed layer results, and momentum is spread deeply, diluted by $1/D$, where D is the layer depth. Both inertial oscillations and lower frequency wind-driven flows are wellreplicated by this model. This model, referred to as the PWP model, has been verified against observations (Figure 5.3.4.2-1) and adapted by the U.S. Navy for predicting the penetration depth of the diurnal mixed layer and sea surface temperatures.

In the absence of time series of the local air-sea heat flux, the climatological mean mixed layer depth for the site should be used. Then, a simple model that predicts the wind-driven motion of that mixed layer as a slab can be constructed of the form:

$$\begin{aligned} \partial u/\partial t - f_c v &= (1/\rho_w D)\tau_x - cu \\ \partial v/\partial t + f_c u &= (1/\rho_w D)\tau_y - cv \end{aligned} \qquad (5.3\text{-}6)$$

where u, v are the east, north velocity components, f_c is the Coriolis parameter, D is the thickness of the slab mixed layer, τ_x, τ_y, are the east, north components of the wind stress, and c is a damping parameter for the inertial oscillations, usually chosen to be about (5 days)$^{-1}$. Pollard (1970) and Weller (1978) give examples of such models.

The wind-driven currents predicted by such models then need to be added to an estimate of the geostrophic flow field. That may also come from a model, though the general circulation models are regional, basin-scale and most often global in order to properly capture the dynamics that control the geostrophic flow field. Figure 5.3.4.2-2 provides an example velocity field from a tropical ocean model run by NCEP. Some caution is recommended in choosing such a model. They do not all attempt to resolve the vertical structure of the velocity field, approximating the ocean as a small number of layers. However, the U.S. Navy and others are running such models with increasing small spatial resolution (down to 1/16th degree of latitude/longitude). Thus, velocity products for a given site or a transit route should be sought from a high resolution, eddy resolving model. The character of the eddies and boundary currents in such models now resembles that which is observed, but it remains beyond the capability of the models to either hindcast or predict in exact detail the mesoscale (100 km) to small scale (10 km) surface current fields. Such high resolution models are being run within the U.S. Navy at the Naval Research Laboratory (NRL) at Stennis Space Center.

In coastal locations or those characterized historically by strong eddy, jet, and frontal variability, it may require that such variability be added by hand to model-produced wind-drive and geostrophic flow fields in order to properly characterize the variability and magnitude of the flows that will be encountered. In coastal locations, tidal flows should also be added, if not included already. Models for prediction of tidal flows are available. In all case, historical surface current information should be used to cross-check the model velocities.

5.4 Statistical Description of Environmental Components

The ultimate objective of this project is to determine appropriate design environmental events and their combinations based on the various performance requirements for the MOB. In view of the large uncertainty associated with the environmental events such as the random occurrence time, random intensity and duration in each occurrence, and possible correlation among different events due to the physical processes that generate the events, a statistical description using probabilistic models is necessary. The event combination and joint probability issue is important since the most severe MOB response (including the response of different MOB structural components) may not occur at the maximum of one component such as wave, but at a particular combination of wind, wave, and current of certain direction. It is presumably conservative to apply the maxima of all events simultaneously in specifying the design event. This is particularly true in view of the large number of performance requirements for the MOB in widely different environments. The description draws from the existing methods and models which have been developed in the past but with consideration of the special problems related to the mission and operation of the MOB such as its size and response characteristics. It is qualitative in nature and references to the literature are given for details.

5.4.1 Random Occurrence of Environmental Events in Time and Space

The MOB in a given operation condition, e.g., in transit, stationary separated, or stationary connected, etc., sees the environmental events, wind, wave, and current as random functions of time and space. The response to the event may be insignificant, quasi-static, or dynamic depending on the amplitude and the time and length scale of fluctuation of the events. Further one could classify the response as transient (from a period of excitation that is short relative to the natural modes of the structure) or stationary (from a loading whose statistics are constant over a period of several natural periods of the structure). If the temporal fluctuation scale is in the range of the MOB natural periods of oscillation and the spatial fluctuation scale in the range of the size of the MOB, the response would be significant and likely to be dynamic in nature, otherwise, quasi-static, or negligible. Therefore, it is convenient to classify the environmental events statistically according to the scale of fluctuation as being either long-term or short-term.

For storm generated events, it is common practice to treat the occurrence of storms over time and space and storm intensity variation as long-term fluctuations. Variations of wind, wave, and current within each occurrence of storm are characterized as being short-term. A schematic sketch of the long-term and short fluctuations in time of the environmental events is given in Figure 5.4.1-1, where occurrence time, duration, and intensity are described by random variables and within-storm fluctuations by random processes. A similar approach may be used for internal waves, i.e. the occurrence of the waves over time and space as being long-term and the variation of currents during each occurrence as being short-term.

5.4.1.1 Long-term Statistical Description

Storms and internal waves may be characterized by their duration, frequency, and intensity at a given location. Table 5.4.1.1-1 shows some typical values of these parameters for a number of events, which may be of interest in design of a MOB. The table values are the mean or maximum value, individual events have duration and intensity variations which can be described by random variables with appropriate probability distributions. The frequency is also the mean value. Individual events may be modeled by appropriate occurrence models. The occurrence in time of an event therefore can be modeled by a random process with a specified mean frequency, a mean duration and an intensity distribution. Such occurrence models allow an easy treatment of the probability of joint occurrence of different events to be given in Section 5.4.2. Since hurricanes and typhoons generally produce the most severe environmental conditions, a more detailed description of these storms is given in the following.

Event	**Characteristic Duration of Intense Phase at Given Site**	**Characteristic Frequency at a Given Site/Year**	**Max. Intensity (upper bound)**
Extra-tropical Cyclones	1 day	70	80 mph (wind speed)
Severe Tropical Cyclones (Hurricane/ Typhoon)	5 hours	0.6	175 mph (wind speed)
Tornadoes	0.5 min	0.005	250 mph (wind speed)
Convective Storms (Thunderstorm)	30 min	20	70 mph (wind speed)
Internal Waves	3 min	Varies greatly	0.5 m/s (current speed)

Table 5.4.1.1-1. Long-term Event Characteristics

A commonly used model for hurricane/typhoon occurrence in time is the simple Poisson process. The only parameter required is the mean occurrence rate. Mean occurrence rate statistics may be collected from historical records for a site in a given reference area or zone. In spite of the rather restrictive assumptions, the simple Poisson process model has proved to be robust compared with occurrence statistics. More refined models may be used such as periodic Poisson and renewal processes. Experience, however, has shown that the improvement in accuracy is quite marginal and generally does not justify the additional effort for the higher moments of occurrence statistics required for these models. The simple Poisson process model, therefore, allows one to evaluate the probability of occurrence of an environment event or a storm-caused severe response for a given period of time. For example, the probability that an event of interest, E, occurs at least once in a time period (0,t) can be given by:

$$P(E, t) = 1 - e^{-\nu P(E) t} \tag{5.4-1}$$

in which ν is the mean storm occurrence rate, P(E) is the probability of occurrence of the event, E, given the occurrence of the storm. For example, E may be a given threshold level of wave height or a MOB response being exceeded. The simple Poisson model also provides a basis for probability evaluation via Monte-Carlo simulation to be mentioned in a later section.

Additional important parameters of a hurricane/typhoon are the intensity, size, translation speed and direction, and storm surge. The commonly used intensity measure is the pressure drop at the center of the storm, ΔP. The size can be measured by the radius from the storm center to the point of maximum wind speed, R_m. One can model ΔP, R_m and the translation speed V_f, by random variables using statistics of historical storms. For example, a joint lognormal distribution has been found to be a good model for hurricanes in the Gulf of Mexico (Wen and Banon (1991)). In other words, the logarithmic transformations of these three parameters follow a joint normal distribution:

$$f(\Delta p, R_m, V_f) = f(\mathbf{x}) = \frac{1}{(2\pi)^{3/2}|C|^{1/2}} e^{-\frac{1}{2}(x-\mu_x)^t C^{-1}(x-\mu_x)} \tag{5.4-2}$$

in which **x** is the vector of the logarithmic transformations of the three parameters, i.e., $x_1 = \ln \Delta P$, $x_2 = \ln R_m$, and $x_3 = \ln V_f$; $\mu_\mathbf{x}$ is the vector of the mean values of **x,** and **C** is the covariance matrix of **x**:

$$\mathbf{C} = \begin{bmatrix} \sigma_{x_1}{}^2 & \rho_{1,2}\sigma_{x_1}\sigma_{x_2} & \rho_{1,2}\sigma_{x_1}\sigma_{x_3} \\ \rho_{1,2}\sigma_{x_1}\sigma_{x_2} & \sigma_{x_2}{}^2 & \rho_{1,2}\sigma_{x_2}\sigma_{x_3} \\ \rho_{1,2}\sigma_{x_1}\sigma_{x_3} & \rho_{1,2}\sigma_{x_2}\sigma_{x_3} & \sigma_{x_3}{}^2 \end{bmatrix} \tag{5.4-3}$$

in which σ is the standard deviation and ρ is the correlation coefficient. The joint probability is thus completely governed by the mean values, standard deviations, and correlation coefficients of these three parameter. The statistics of the three parameters for the hurricanes in the Gulf of Mexico are shown in Table 5.4.1.1-2. The prominent features are the large coefficients of variation of the storm parameters and the negative correlation between ΔP and R_m, and ΔP and V_f. Although the correlations are small, they may have an impact on the MOB response. The direction of the storm track is also an important parameter since it may have an impact on the MOB response depending on the angle between the storm track and the MOB orientation. It can be modeled by a random variable. For example, for GOM hurricanes, a beta random variate for direction distribution independent of the rest of the storm parameters is found to be a satisfactory model (Wen and Banon (1991)). The storm surge can be also modeled as a random variable but generally strongly dependent on the storm intensity parameter.

Parameters	Mean Value	Coefficient of Variation
ΔP (mb)	45.04	0.507
R_m (nautical miles)	24.94	0.433
V_f (knots)	9.72	0.361

Table 5.4.1.1-2. Statistics of GOM Hurricanes Based on 1900-1983 Data

$\rho\ (\Delta P, R_m) = -0.351$

$\rho\ (\Delta P, V_f) = 0.260$

$\rho\ (R_m, Vf) = -0.108$

Once these storm parameters were specified, the wind, wave, and current fields of the hurricane/typhoon can be determined using existing parametric or numerical field models (e.g., Cooper (1988)). That is, these models allow one to monitor the change in wind speed and direction, significant wave height, H_S, and direction, θ_V, current speed, V_C, and direction, θ_C as well as storm surge, Z_{SU} at a given site relative to the storm center during the passage of a storm. Figure 5.4.1.1-1 shows an example of the wave and current fields based on the foregoing models. Any correlations between wind, wave, and current are included in the field model. The advantage of such modeling is that it allows direct calculation of the MOB response time history during a passage of a storm and calculation of response statistics via a Monte-Carlo method (Wen and Banon (1991)). More details will be given Section 5.4.3.

It is pointed out that parametric models such as the joint lognormal distribution proposed above need to be verified against records and hindcast data based on a statistical goodness-of-fit test. A satisfactory parametric model is not always easy to establish. In case of sufficient amount of historical data, nonparametric, empirical joint distribution models can be also used for the same purpose. In this case, the multi-dimensional empirical distribution (e.g., based on a multi-dimensional histogram) can be used in place of Eq. 5.4-2. A note of caution is necessary here, however, that in order to have a reliable empirical model which is not unduly constrained by a small number of historical observations, a large sample of data is necessary.

The probabilistic models and statistics of the storm parameters allow one to evaluate the conditional probability of a given event of interest, P(E), given the occurrence of a storm. The procedure and methodology for evaluation of P(E) will be given in Section 5.4.3.

Although the brief description given above is for hurricanes/typhoons, largely based on past experience with platforms in GOM, a similar treatment of events such as internal waves is possible.

Alternatively, since wave loads generally dominate the MOB response, the storm can be characterized by the sea state in terms of significant wave height, H_S. For example, Banon et al. (1994) modeled hurricanes with H_S > 8 meters by a Poisson process in time with a mean occurrence rate, ν, and a three-parameter Weibull distribution for H_a given the occurrence of a hurricane:

$$P(H_S < h) = 1 - e^{-\left(\frac{h-\varepsilon}{h_0-\varepsilon}\right)^k} \qquad (5.4\text{-}4)$$

in which ε is the lower limit (8 meter), h_0 and k are the scale and shape parameters respectively. ν, h_0, and k can be determined from hurricane and wave statistics. Other storm parameters include wave direction θ_V, peak period of wave spectrum T_p , wind speed V_W and direction θ_W, current speed V_C and direction θ_C, can be modeled as normal random variables with correlated mean values and constant coefficients of variation. The correlations are introduced via simple functional (linear or power) relationships among the parameters from regression analysis of metocean data (e.g., Forristall et al. (1991), Banon et al. (1994)). For example, the following mean value equations have been proposed in Banon et al. (1994):

$$\begin{aligned}
&T_p: \text{ normal with mean} = a_1 H_S^{\,b} \\
&V_W: \text{ normal with mean} = a_1 + a_2 H_S \\
&V_C: \text{ normal with mean} = a_1 + a_2 H_S + a_3 V_W \\
&\theta_V: \text{ normal with constant mean} \qquad (5.4\text{-}5) \\
&\theta_W: \text{ normal with mean} = a_1 + a_2\,\theta_V \\
&\theta_C: \text{ normal with mean} = a_1 + a_2\,\theta_V
\end{aligned}$$

All above random variables have constant coefficients of variation independent of other storm parameters. These random variables are, however, correlated through these functional relationships among the mean values. This method of modeling allows one to evaluate statistics of structural response using simplified response models and random vibration analysis (e.g., Banon et al. (1994)).

The joint probability method has been proposed for risk studies of hurricane damage. The method fundamentally depends on the modeling of the joint probabilities for the hurricane or typhoon characterizing parameters, or alternately assuming the properties to be uncorrelated. The large data base from the National Weather Service on Guam, providing many years of typhoon characteristics in the general region of Guam, enables these assumptions to be examined. Three of the more significant parameters are the deviation of the central pressure from the regional pressure (DP), the maximum wind speed (SPD), and the radius to maximum wind (RAD). A plot of DP versus RAD is given in Figure 5.4.1.1-2 for 372 events. The data generally lies within a triangle, with very large DP and very large RAD never jointly occurring. Some sort of joint exponential probability law might appear to work for this pattern, but the pattern is too "triangular" and statistical tests do not favor that. A graph of DP versus SPD is shown in Figure 5.4.1.1-3. Here, a sort of elliptical pattern is demonstrated, with the greatest density of data occurring non-centrally in the lower left of the scatter plot. It has been suggested that a bivariate lognormal probability law might work for this data. A LogLog plot of this same information is given in Figure 5.4.1.1-4. The Log-Log assumption is appealing, but the greatest density is not centered in the cloud of points. At any rate there are clearly problems with modeling the joint probabilities.

Similar data from the Gulf of Mexico and Atlantic Ocean is much more sparse, and limited to more severe storms. A plot of central pressure versus storm radius is presented in Figure 5.4.1.1-5. The same general triangle of concentration is present although there is really not enough data graphed to allow very strong conclusions.

The severe difficulty of using the joint probability method for high dimensional multivariate problems led to the consideration of alternatives. Over the past decade a method called "empirical simulation techniques" or EST has been developed which proceeds directly from the empirical data to the risk computations without a required parametric modeling effort. These techniques appear to offer benefits in the MOB design process. However, rather extensive databases are required from actual measurements or hindcasts.

Whatever methods are used in the evaluation of MOB designs, some type of substantial data are required. It appears important to the effort for ONR to obtain access to as many industry and national databases as possible. Several Industry data compilations are:

(1) The "Gumshoe" study coordinated by Chevron,

(2) The older "ODGP" project organized by the Shell Development Company, and

(3) The "WADIC" data measurements in the North Sea collected by The Continental Shelf Institute (Norway).

There are others, of course, but the point is that detailed and accurate design starts with reliable input data and extends for there to consider realistic linear and nonlinear attributes of waves, tides, winds, currents, and other oceanic processes. It is particularly important to include the real frequency and directional variability present in the actual oceanic chaos, and not an over simplified and regularized version of the phenomena.

5.4.1.2 Short-term Statistical Description

Short-term fluctuations include those in wind, wave, and current over one to three hours during which the storm intensity or sea state remains approximately constant. One can describe the random fluctuation of these events by a multi-dimensional stationary random process (or random field). The time domain fluctuation of such a process can be characterized by a power spectral density function. The spatial correlation can be described by coherence function for wind turbulence and directional spectrum for waves spreading as described in previous section on winds and waves. The Gaussian process is commonly used model for wind turbulence and wind-generated linear waves. The short term fluctuation therefore can be characterized by the parameters of a stationary Gaussian process.

Short-term statistics of interest in response analysis of the MOB include probability distribution and parameter of wave height, period and direction, turbulence intensity as function of height, scale of fluctuation of turbulence along wind and across wind, current strength and direction and variation with depth. All these random processes and statistics are conditional on a given set of storm or sea state parameters. The fluctuation of wind turbulence and current variation have been discussed in Sections 5.2.1 and 5.2.2. The amplitude of a narrow-band Gaussian process such as wind-driven linear waves follows a Rayleigh distribution. Empirical evidence indicates that individual wave heights H in a given sea state H_S can be modeled by a two-parameter Weibull distribution (Ward (1978));

$$F_H(H) = 1 - e^{-a\left(\frac{4H}{H_s}\right)^k} \tag{5.4-6}$$

in which a and k are constants determined from wave statistics. The above distribution can be interpreted as a more general form of the Rayleigh distribution to accommodate the nonlinear wave behavior. The maximum wave height, H_{max}, over the given sea state duration can be shown to follow an Extreme Value Type I (Gumbel) distribution with a small coefficient of variation (< 10 %). In view of the large uncertainty in the long-term parameters, the small uncertainty in H_{max}, is generally ignored and the maximum wave height is given approximated by:

$$H_{max} = 1.8\ H_S \tag{5.4-7}$$

which is close to the mean value of H_{max}. Since structural response is sensitive to the wave period, the joint distribution of wave height H and wave period T is of interest. For a Gaussian wave process, the joint distribution can be shown to be given by the following (Longuet-Higgins (1975)):

$$P(h,z) = \frac{h^2}{\sqrt{2p}} e^{-\frac{1}{2}h^2(1+z^2)} \tag{5.4-8}$$

in which

$h = H/\sqrt{m_0}$, the non-dimensional wave height

$\zeta = (T - \overline{T})/\nu$, the non-dimensional, zero-mean, wave period

$\overline{T} =$ the mean wave period

$\nu = (m_2/m_0)^{1/2}\overline{T}$, the wave spectral width factor

m_0 = zero-th moment of the wave spectrum, free surface variance

m_2 = second moment of the wave spectrum.

A plot of this joint distribution is shown in Figure 5.2.3.5-3.

Again, empirical distributions based on observational data show slight variations.

The joint distribution of wave height and period is of special importance since maximum structural response may occur at a period which does not corresponds to the maximum wave height.

5.4.2 Joint Occurrence of Environmental Events

An important consideration in response analysis and design is the joint occurrence of severe environmental events, which can cause the MOB failure to meet the operational requirements. Of primary concern are:

(1) simultaneous occurrence of severe events most damaging to a MOB; generally these have a very small probability of occurrence

(2) wind, wave, and surface wind-driven current are highly correlated in intensity and direction in a given storm

(3) since there are a large number of MOB operational requirements, some or all of them may not be controlled by a single most severe event such as wave height, rather by a combination of wind, wave (height and period) and current, each of a particular direction.

The last issue is MOB response dependent and will be addressed in Section 5.4.3. This section will be devoted to the first two issues and their implications in MOB design.

In view of the large uncertainties associated with all environmental events, the operational requirements may be realistically stated only in probabilistic terms. In other words, the satisfactory performance of MOB in different operational conditions should be stated in terms of allowable probabilities (or frequencies) of various limit states associated with each operation condition.

The joint occurrence of severe environmental events needs not be considered if the probability of such joint occurrence is below the allowable probability. In other words, the requirements are automatically satisfied without even considering the MOB response. For instance, if the return period for one type of event is 20 years, and another uncorrelated event is also 20 years, then the return period of the two occurring simultaneously is much less than 100 years, and so the combination need not be considered. This elimination process may prove to be useful in narrowing down a few critical combinations among a large number of potential joint events. The method of modeling of the long-term fluctuation in time as described in the previous section can expedite the evaluation of the joint occurrence probability as follows:

$$P(E,\, t) \approx 1 - e^{-\nu_E t} \approx \nu_E t \qquad (5.4\text{-}9)$$

in which

E= the joint event of interest,

t = the time period of interest (a year, or given period of certain operation condition).

ν_E = the mean occurrence rate of the event E.

An underlying Poisson process has been assumed for joint occurrence. ν_E is the only parameter required and it is a function of the mean occurrence rates and mean durations of the individual events. If the occurrences of the events are statistically independent, it can be shown that for combination of two events (Wen (1990)):

$$\nu_E = \nu_1 \nu_2 (\mu_{D_1} + \mu_{D_2}) \qquad (5.4\text{-}10)$$

in which ν_i and μ_{D_i} are the mean occurrence rate (frequency) and mean duration of the individual events. Note that the information of the duration distribution is not required. The results given above have been shown to be insensitive to duration distribution assumption.

As an illustrative example, assume that at a given site, the hurricane has an occurrence rate of 0.5 per year and a mean duration of intense phase of 5 hours, and a major ocean front occurs at a rate 0.1 per year with a mean duration of 1 hour. The above equation gives a mean joint occurrence rate of these two events:

$$\nu_E = 0.5 \times 0.1 \times \left(\frac{5+1}{24 \times 365}\right) = 3.42 \times 10^{-5} \text{ per year} \qquad (5.4\text{-}11)$$

Therefore, it is clear that the probability of joint occurrence of these two events based on Eq. 5.4-9 would be very small. If it is lower than the allowable annual probability of a limit state that may be caused by the joint occurrence of these two events, then the joint occurrence can be eliminated from the structural response and design consideration. For joint occurrence of three independent events, the joint occurrence rate can be shown to be:

$$\nu_E = \nu_1 \nu_2 \nu_3 (\mu_{D_1}\mu_{D_2} + \mu_{D_2}\mu_{D_3} + \mu_{D_1}\mu_{D_3}) \qquad (5.4\text{-}12)$$

The general trend is that if the occurrences of the individual events are statistically independent, the more events one considers in the combination, the smaller the joint occurrence rate. The situation changes, of course, when there is high degree of correlation in occurrences among the individual events. The above solution can be extended to the cases where the individual events are correlated in occurrence time, (clustering of occurrences within each non-Poissonian event, and between events), duration and intensity. Details can be found in Wen (1990).

Another type of joint occurrence of interest is that of wind, wave, and current of certain magnitude and direction produced by storms. As mentioned in the foregoing, these events are highly correlated because of the physical process of a storm. According to the methods of modeling given in the previous section, the joint occurrence can be considered using either an empirical field model through simulation (to be given in more details in Section 5.4.3) or directly using a parametric joint distribution model such as suggested in Eqs. 5.4-4 and 5.4-5. As an illustrative example, consider a critical event of joint occurrence of significant wave height exceeding a threshold level in a certain direction sector and current exceeding another threshold level in a different direction sector. One can evaluate the probability of this joint event given the occurrence of a hurricane by conditioning on significant wave height and direction as follows:

$$P(E) = P[(H_s > h_0) \cap (\theta_0 < \theta_v < \theta_0 + \Delta\theta); (V_c > v_0) \cap (\theta_1 < \theta_c < \theta_1 + \Delta\theta)]$$

$$= \int_{h_0}^{\infty} P(V_c > v_0 | H_s = h) f_{H_s}(h) dh \int_{\theta_0}^{\theta_0+\Delta\theta} P(\theta_1 < \theta_c < \theta_1 + \Delta\theta | \theta_v = \theta) f_{\theta_c}(\theta) d\theta_c \qquad (5.4\text{-}13)$$

in which the conditional probability and probability density functions are obtained from Eqs 5.4-4 and 5.4-5. The probability of such joint event over an operation period, t, of the MOB can be then calculated from:

$$P(E, t) = 1 - e^{-\upsilon P(E) t} \approx \upsilon P(E) t \qquad (5.4\text{-}14)$$

in which u is the mean occurrence rate of the storm with $H_S > 8$ meters. Again, if this probability is lower than the allowable limit state under this particular combination of event, then the joint occurrence can be eliminated from structural response evaluation and design consideration.

5.4.3 Long-term Statistics of Structure Response

The performance requirement of a MOB in a given operational condition can be stated in terms of a particular MOB response not exceeding a critical threshold level, or a limit state. For example, the limit state could be in terms of a certain motion critical for MOB operation, or in terms of stress in a certain structural component critical for MOB survival in severe environmental conditions. In other words, the critical event and event combinations for design are MOB response dependent. In view of the large uncertainty in environmental events, the MOB performance can not be enforced in absolute terms such as not exceeding the limit state under the maximum possible environmental conditions. Such an approach could lead to design criteria which are unduly conservative and economically unfeasible. Satisfactory performance can be realistically ensured only in terms of the MOB limit state probability over a given period of time (or alternatively, frequency of occurrence) below an allowable (or target) value. In the lifetime of a MOB, a large number of limit states need to be considered corresponding to different operation conditions. The target values may be established from the operational requirements and economical constraints (such as costs). The objective of design is then to select a set of design cases, (environmental events and event combinations, in combination with the condition (damaged or intact) and the project phase (disconnected, connected, transit etc.)) for the design of the MOB such that the limit state probabilities for each operation condition will be lower than the corresponding target values.

To achieve the above design objective a methodology is needed for evaluation of long-term MOB response statistics and probability and developing criteria for determining the design events and event combinations. Drawing on past experience with offshore platforms, the response fluctuations and probability can be described again in terms of being short-term and long-term, similar to the method used in describing the environmental events. Two approaches can be used.

5.4.3.1 Analytical Method

Using the short-term random process and long-term random variable models for the environmental events as input, the response of the structure can be treated likewise as a random process and random variables as output. That is, the response time history can be modeled by a random process and the response of interest such as maximum deformation as a random variable. The problem again can be conveniently described as that of short-term and long-term (Banon et al. (1994)).

Short-term Problem

The short-term problem concentrates on the probability and statistics of structural response during a given storm. Provided the structural response is linear with respect to wave height, the response during the most intense phase of the storm can be modeled by a stationary random (Gaussian) process. The important probabilistic characteristics of the process such as root mean square (RMS) value, zero-crossing period, etc can be modeled as simple functions of the long-term storm or event parameters such as H_s, T_p, wind speed, current speed, etc. These input-output relationships can be established by regression analyses of a large number of structural model responses under hindcast of historical storms. A response surface method may be used to obtain the simple input-output relationship in polynomial form. The event of interest such as the occurrence of a limit state can be described in terms of the outcrossing of the response random process of a given threshold level. The probability of such an occurrence given that of a storm is then given by:

$$P(E|\text{storm}) = 1 - e^{-\nu(E)t_S} \tag{5.4-15}$$

in which a Poisson outcrossing process is assumed

E = the event of a given limit state being exceeded

ν = the mean occurrence rate of E from a outcrossing analysis

t_S = storm duration.

To calculate ν, a First Order Reliability Method (FORM) or Second Order Reliability Method (SORM) can be used. Computer software of FORM/SORM (Liu et al. (1989)) is available for this purpose.

Long-term Problem

The short-term solution of the response probability given in the foregoing is conditional on a given set of storm parameters. To incorporate the variability of storms, one needs to consider the storm occurrence rate and distribution of other random storm parameters. The unconditional limit state probability over a given time period (0, t) is then given by:

$$P(E, t) = 1 - e^{-\nu_S E_S[P(E|\text{Storm})]t} \tag{5.4-16}$$

in which ν_S = mean storm occurrence rate; E_S [ ] = operation of expectation with respect to random storm parameters. Again, a computational efficient method based on FORM/SORM can be used for evaluation of the expectation (Wen and Chen (1989)) when the number of parameters is large.

5.4.3.2 Time History/Monte-Carlo Method

A more direct method for evaluating structural response statistics and probability is by simulating the environmental events and exciting force time histories on a computer and solving for the structural responses. This process is repeated for a large number of times reflecting the statistical variation of the environmental events and the structural responses. From this ensemble of response time histories, one can calculate the required response statistics and probability. Again, it is convenient and computationally efficient to separate the calculation into long-term and short-term response simulations (Wen and Banon (1991,1995)).

Short-term Simulation and Response Calculation

For a given set of the long-term parameters, the dynamic structural response depends on the short term fluctuations such as wind profile and turbulence, wave amplitude and period, and current profile. A structural response model is needed which relates the response of interest to the fluctuation parameters. If a detailed time history response method is used, one needs to simulate the short term fluctuation using the wind and wave spectra, spreading and coherence functions. The computation required is generally extensive, and probably prohibitive. Alternatively, one can use simple empirical models, which give the structural response as function of the long-term parameters. For example, the response of interest such as a displacement with a certain probability of exceedance during the most intense phase of the storm can be expressed as simple linear or nonlinear (polynomial or power) functions of the wave height and period with the same probability of exceedance, and associated wind and current speeds in the same direction of the wave. The empirical formula are determined from a very large number of response analyses of the structure under various environmental conditions. The bias and uncertainty of such empirical equation need to be established and incorporated in the long-term response statistics and probability. The latter approach has been most popular in the evaluation of response probability of offshore platforms. The response of the structure obtained is then one sample during the passage of one simulated storm. A large number of storms with various combinations of long-term parameters are simulated, each produces a set responses of interest for evaluation or design.

Long-term Simulation

This is simulation of passage of a storm or an internal wave at the site. Refer to long-term characterization of environmental events in Section 5.4.1. These events are characterized by their model of occurrence in time (e.g., Poisson process), storm track direction (e.g., beta distribution), storm intensity, size, and translation speed (e.g., jointly lognormal random variables), and storm surge (random variable). These random parameters are generated on computer and from which one can construct the wind, wave, and current fields from the parametric or numerical models (Figure 5.4.1.1-1). In generating the storm parameters, one can use the parametric distribution models as mentioned. If such models are difficult to establish and sufficient storm data are available from records or hindcasts, one can use a non-parametric procedure based on the histogram or multi-dimensional histogram for the same purpose. Referring to Figure 5.4.3.2-1, the wind, wave, current, and surge time histories at the site during a storm passage therefore can be calculated and from which one can evaluate the maximum structural response of interest using an appropriate response model or computer software as mentioned in the previous section. Figure 5.4.3.2-2 shows an example of a tension leg platform response time history during the passage of a hurricane (Wen (1992)). The platform is located at a distance of one half of the radius of the maximum wind speed and to the right of the storm track. The horizontal axis is the relative position in meters of the storm center to the platform site, along the storm track. The process is repeated for a large number of times. The large sample of the response provide the data base to calculate the long-term response statistics and the probability of the response over a specified time period.

5.5 Fatigue

Fatigue damage in members is caused by experiencing large numbers of stress cycles at levels significantly below design stress levels required for strength. During a 40-year life, a MOB may see about 1×10^8-wave cycles (based on 10-sec wave period). Stresses of only about 5 ksi, with this number of cycles, would result in fatigue being as important as strength in the design.

In some marine environments fatigue considerations govern the design of marine structures, and this is expected to be the case for the MOB.

Fatigue is particularly important when the structure is subjected to similar wave environmental conditions over its whole life, rather than a very few isolated extreme storms. This is the case in the North Atlantic deployment site, where the MOB is to be designed to spend a significant fraction of its life. In this region, there is continuous exposure to quite large storms throughout the year, and so the MOB will experience a relatively large number of cycles of loading with loads not greatly less than the peak loads from the largest expected waves.

Fatigue frequently dictates member sizes and connection details for offshore platforms in the North Sea, and large floating structures will be even more susceptible to fatigue. This is because, for large-sized members, since wave forces depend on inertial effects in the waves, these forces depend linearly on wave heights, while wave forces on offshore platforms vary with the square of the wave heights, since these forces are mainly from drag. As a result, smaller waves are relatively more important for the MOB than for typical offshore platforms, and we can expect fatigue to be a major effect in determining the steel thicknesses and connection details in the final MOB design.

This situation is in contrast to the Western Pacific region, where the commonly-occurring storms are far smaller than the typhoons that occur quite rarely at any one site. Under these conditions, the structure is designed for the typhoon loads, and the stresses during the remainder of the life of the structure are relatively much less, and therefore contribute less to the total fatigue damage.

During transit, there will be potential for accumulation of fatigue damage, but since it is only in this phase about 5% of its life, and in wave environments far less severe than the North Atlantic, although it should not be neglected, it is not likely that fatigue will be of much importance during this time, unless there are some parts of the structure that are peculiarly sensitive to the transit conditions - the forward motion, or the lightship condition.

The procedures for estimating fatigue damage are well established for both fixed and floating structures, and need not be detailed here. The principal area where the MOB differs from other marine structures is its great length, and so a few comments will be made about the effect of this on the fatigue calculations. The issue to be addressed here is how do the uncertainties in wave coherence from point to point in the MOB relate to fatigue damage computations.

Since fatigue damage is caused by commonly-occurring waves, rather than the extreme waves from rarely-occurring storms, the issue of wave coherence could be more important for fatigue computations than for extreme seas. Because the waves are smaller than the extreme waves, they will generally have shorter wavelengths. Thus the MOB will span a greater number of wavelengths and so the uncertainty in wave coherence effects will be greater than for extreme seas.

In the North Atlantic, the most commonly occurring seastates have zero-crossing wave periods rather less than 8 sec, although damage may be more important from the larger and hence longer-period waves. Considering, therefore, waves of period 10 seconds, we see that their wavelengths of 500 ft are only 1/10 of the length of the MOB, and about 1/2 of the length of a module. In the connected conditions, stresses from waves are expected to be high in the modules, and so will contribute a major part of the fatigue damage to many parts of the structure. So there is a measure of uncertainty that arises in the fatigue calculations from the uncertainties in wave coherence.

Unfortunately, the estimated annual fatigue damage varies as the third or fourth power of the stress levels, so uncertainties in the computed stresses are translated into greater uncertainties in the estimated fatigue damage.

Environmental statistics for strength design need to be gathered from data describing large, rather infrequent storms, so in the Western Pacific, it may be appropriate to collect hurricane information. However, for fatigue computations, it is the commonly-occurring waves that build up the fatigue damage. Therefore, the emphasis for fatigue computations is on a complete picture of the environment, from the relatively small seastates that are exceeded for large fractions of the year, through to the larger storms that occur annually or less. In many areas of the world these data, collected routinely by ships and other agencies, are available for many decades, giving a substantial data base for these calculations.

The conclusions to be drawn from the above remarks are:

1. We can expect fatigue to be a major player in the design of MOB members and connections, and should be considered from the very beginning, for all parts subjected to stresses from waves. This will include both structural members and connectors.
2. Since the MOB may be designed to spend a significant fraction of its life in the North Atlantic, we can expect this region to contribute almost all of the fatigue damage that the structure will see during its life.
3. A large amount of the fatigue damage is expected to accumulate in the connected phase, since the MOB will probably spend most of its time connected, except when riding out extreme storms, and the stresses in many parts of the structure are higher when connected, than when disconnected.
4. Again, since most of the fatigue damage is expected to accumulate in the connected phase, the uncertainties of wave coherence, of both broadside and head seas, may play a significant role in the uncertainty in the estimated fatigue damage.

Environmental data for fatigue design is to concentrate on non-extreme conditions, whereas that for strength design concentrates on rarely-occurring storm events.